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Structural, stratigraphic and sedimentological characterisation of a wide rift system: The Triassic rift system of the Central Atlantic Domain

Sophie Leleu (1)

Adrian J. Hartley (2), Cees van Oosterhout (3), Lorcan Kennan (3), Katrin Ruckwied (3), Keith Gerdes (3)

(1) corresponding author : sophie.leleu@bordeaux-inp.fr; phone: 00 33 (0) 5 56 84 69 29

Bordeaux-INP/ ENSEGID, EA4592, Université Bordeaux Montaigne, Pessac, 33607, France

(2) University of Aberdeen, School of Geosciences, Aberdeen, AB24 3UE, UK

(3) Shell, E&P, Rijswijk, The Netherlands

1. Introduction

In the Central Atlantic domain, although much work has been published over the last 40 years (Baudon et al., 2012; Manspeizer 1988a, b; Olsen, 1997; Piqué & Laville, 1995; Withjack et al., 1998) Triassic rifting remains poorly understood. Uncertainties associated with Triassic rift systems are due to a range of characteristics which include: 1) a lack of good quality offshore data, partly because the Triassic section is deeply buried and the available data are old, 2) limited outcrop information including lack of recognition of major unconformities that record complex basin fill histories, 3) lack of a robust stratigraphic framework due to mostly barren continental deposits with rare age-diagnostic fossils, and 4) scattered available data with few penetrations by exploration wells particularly along the eastern coast of North America. Consequently, Triassic rifting is often considered to be the “poorly constrained” event between Late Palaeozoic Caledonian/Variscan collapse and Early Jurassic sea-floor spreading in the Atlantic Ocean. It is crucial to improve our understanding of Triassic rift processes and basin-fill architecture to constrain the nature of crustal thinning prior to Atlantic sea floor spreading. In this review, the rifting stage that initiated Pangea breakup (Frizon de Lamotte, 2015; Olsen, 1997) can be considered as a wide rift system with complex and variable geometries developed at a continental scale within which syn-rift deposits remain continental for over 35 My. The Triassic basin architectures show that uplift occurs during rifting and that the upper crust geometry records deep lithospheric and/ or mantle dynamics.

Seventeen Triassic basins have been studied along the Central and North Atlantic between Florida and Newfoundland on the western Atlantic margin and between Morocco and the Bay of Biscay on the eastern margin of the Atlantic. Additional basins may be present offshore of the African craton south of Morocco, but we are not aware of any public data available for these areas. This work documents the Triassic/ early most Jurassic sedimentary fill, stratigraphy and basin architecture for basins in both North America and the conjugate margin of NW Africa. Data were compiled and synthesized from both margins of the Atlantic. The types of data and methods used by previous authors are very diverse and therefore the amount and quality of data vary from basin to basin. We aim to present consistent stratigraphic nomenclature and facies types when comparing and contrasting different basins. In this regional scale overview we describe a range of different basin geometries and structural zones, we characterized the main depositional environments and provide stratigraphic compilations that constrain palaeogeographic maps for Anisian, Carnian and Rhaetian times. We discuss the mechanisms that controlled the initiation and long-lived development of the Central Atlantic rift system prior to breakup. Finally, we highlight the key tectono-stratigraphic characteristics of a wide rift system which can be used for comparison with other similar rift systems.

2. Geological setting

The Triassic basins along the North American (NA) Atlantic margin are known as the Newark Supergroup basins (Olsen, 1978). Over 11 basins (**Fig. 1**) have been recognized onshore from Florida (USA) to Nova Scotia (Canada). Over 17 basins are located offshore from Virginia to Newfoundland but considerable uncertainties exist regarding the age and nature of their sedimentary fill. Basins from offshore Nova Scotia and Newfoundland are by far the best know from the offshore area due to extensive hydrocarbon exploration. Basins in Morocco are laterally equivalent to the Nova Scotian basins whereas basins in western Iberia are laterally equivalent to the Newfoundland basins (**Fig. 1A**).

The passive margin of North America has been extensively described and discussed by Withjack, Schlische and Olsen in a variety of papers (e.g. Olsen & Schlische, 1990, 1996; Olsen, 1997; Olsen et al., 2000; Schlische et al., 1990; Schlische, 1993; Withjack et al., 1995, 1998). The NA margin developed in two distinctive stages: rifting and drifting. From south to north, although dating is an issue, some data support the view that drift initiation varied along the margin (Withjack et al., 1998) although alternative interpretations suggest that sea floor spreading occurred synchronously at 195 Ma (Labais et al., 2010; Sahabi et al., 2004;). Most workers dated the rifting of NA basins between Mid Triassic and early Jurassic (e.g. Manspeizer & Cousminer, 1988; Withjack et al., 1998). Withjack & Schlische (2005) summarized the diachroneity of the age of syn-rift phase and age of sea floor spreading along the NA margin, and they defined groups of basins and refer to them as, from south to north (**Fig. 1B**) namely, Southern Segment (from basins in Florida to Taylorsville Basin), Central Segment (From Culpeper to Orpheus Basin), and Northern segment (basins offshore Newfoundland). We use this terminology mainly in a geographical sense for simplicity and also include the conjugate margins, such as the Moroccan basins which are included in the Central Segment and the Portuguese basins in the Northern Segment.

2.1 Inheritance

The pre-Triassic structural framework resulted from amalgamation of different crustal blocks during the late Ordovician/ early Silurian Caledonian Orogeny and the late Carboniferous Variscan Orogeny (Doré et al., 1997; Matte, 2001; Stampfli & Borel, 2002). In addition a poorly constrained period of rifting and thermal subsidence occurred during the Permian in some parts of the study area. The influence of the pre-exiting structural framework and Permian basin development on Triassic rifting and basin development has yet to be fully evaluated but broad trends in basement structure can be defined (**Fig. 1B**). Closure of the Iapetus Ocean and development of the Caledonian Orogeny resulted in a broad suture zone including what is now the east coast of Greenland, NW Scotland, western Norway and the eastern seaboard of North America. The southern and western part of this area was subsequently structurally reworked in the late Carboniferous during development of the Appalachian-Variscan Orogeny (Doré et al., 1997; Matte, 2001; Michard et al., 2010; Rast, 1988; Secor et al., 1986; Simancas et al., 2005; Stampfli & Borel, 2002; Ziegler, 1990). This generated a broad zone of deformation extending across Central Europe through the UK, northern and western Iberia, Morocco and into the eastern seaboard of North America. Both the Caledonian and Variscan orogenies imparted a NNE-SSW to NE-SW trending structural grain along which the Triassic rift basins of the Central and northern Atlantic subsequently developed.

The Permian history of the region is poorly constrained as little or no data are available either onshore or offshore. Where Permian sediments are thought to be present, age constraints are poor and the distinction between Permian and Triassic strata is often difficult due to a lack of biostratigraphic data. In the subsurface Permian strata have not been clearly identified. In areas where Permian sediments have been identified (e.g. Argana Basin Morocco, New Brunswick, Canada) they are thought to represent small rift-basin fills (Hafid et al., 2000; Olsen et al., 2000). It is likely that any phase of Permian basin development was strongly influenced by the pre-existing Caledonian and Variscan structural framework.

2.2 Triassic rifting

The most widely accepted idea is that deposition during the Triassic was synchronous with extensional tectonic activity (Brown, 1980; Leroy & Piqué, 2001; Medina, 1991, 1995; Olsen et al., 1997; Piqué & Laville, 1995; Piqué et al., 1998; Tixeront, 1973; Withjack et al., 1998). Rifting is considered to have been synchronous within the Central Atlantic domain (Piqué & Laville, 1995). At a large-scale, many authors believe that the opening of the Atlantic domain was asymmetric with a half-graben geometry dominating the North American margin and more symmetric grabens along the Moroccan margin (eg. Leroy & Piqué, 2001; Lister et al., 1991; Medina, 1995). In North America, the Triassic basins formed within the Appalachian orogen, and some Palaeozoic reverse faults were inverted during the extensional phase.

Triassic basins have half-graben geometries and are bounded by border fault system (**Fig. 1B**) often inherited from the previous Appalachian orogenic architecture (e.g. Lindholm, 1978a, b; Ratcliffe et al., 1986; Swanson, 1986). Thus most basins trend N-NE to NE. Trans-tensional basins were created where pre-existing faults were obliquely oriented to the extension, e.g. Minas sub-basin (Schlische, 1993). Border faults dip either to the W or to the E and often show a straight and high-angle geometry but some basins have listric or low-angle border faults which re-activated pre-existing thrust faults (i.e. Fundy border fault; Schlische, 1993). Similar basin trends and extension directions have been described in Morocco, where Triassic basins were interpreted to have been formed by transtensional syn-sedimentary tectonics (Piqué et al., 1998) and typically trend NE-SW, following Variscan structures (Piqué & Laville, 1995). Piqué & Laville (1995) regarded Triassic rifting in Morocco as the onset of Atlantic rifting. In Morocco, Leroy & Piqué (2001) suggested that Triassic rifting propagated from E to W and from S to N which is opposite to the thickness trends present in basins and the dating presented by El Arabi et al. (2006). A recent study (Ouarhache et al., 2012) suggested that the easternmost basins of Morocco recorded a Tethyan rifting phase that does not exist in the Atlantic margin (post-dating the 201 Ma CAMP event).

Most of the Triassic basin fill successions in the North and Central Atlantic domain display a similar sedimentological evolution and are dominated by continental deposits (Leleu & Hartley, 2010). The basin-fill sequence commences with a lower sand-prone phase characterised by fluvial, alluvial-fan and fluvio-aeolian sediments. This sequence is overlain by an upper fine-grained, mud-prone unit which may include evaporites and is interpreted as representing deposition in playa/ lacustrine environments and may be interbedded with mafic volcanic sequences. However some southern USA basins vary from this general trend and show two fining upward sedimentary sequences which were recognised by Lambiase (1990). He suggested that each sequence represents a pulse in tectonic activity. Although the sedimentological evolution of the basins studied is similar, the onset of basin filling is strongly diachronous, and the timing and duration of the phases of fluvial and playa deposition vary from one basin to another. In latest Triassic times, the basins in the Nova Scotian/ Moroccan domain recorded the earliest marine transgression into the rift system from the north and east (Dercourt et al., 2000) whilst the southern basins remained in a continental setting. Most of the southern basins were subject to uplift and erosion in the late Triassic/ early Jurassic as indicated by inversion structures that are cross-cut by dykes of latest Triassic age associated with the CAMP event (Withjack et al., 1998).

2.3 Central Atlantic Magmatic Province (CAMP) event

The CAMP event is the largest documented continental large igneous province (LIP) covering millions of square kilometers across four continents (e.g. McHone, 2000, 2003). The CAMP event is considered largely isochronous occurring in less than 840 ± 60ky (Schaller et al., 2012) or 610 ky (Blackburn et al., 2013) and produced great volumes of mafic volcanic successions, sills and dykes used for correlation across the Central Atlantic domain. Indeed the CAMP igneous units are widespread and can be found from South to North America as well as North Africa and South-West Europe. The CAMP volcanics comprise three different lava units in the Fundy Basin (Kontak, 2008) and two main units in Argana (Morocco; El Hachimi et al., 2011) made of complex lava flow lobes but in most places the CAMP is recorded by sills and dykes due to erosion of the shallower sections.

The CAMP had been interpreted as representing a marker unit lying above the Triassic/ Jurassic boundary (e.g. Olsen et al., 1997) but in the revised stratigraphic framework (Cirilli et al., 2009; Dal Corso et al., 2014; Deenen et al., 2011; Ogg et al., 2008;), emplacement of the CAMP volcanics and intrusives occurred during the Rhaetian. Most CAMP units are sills and not extrusive bodies and it is important to recognize this when interpreting ages of sediments around CAMP sills which can form apparently conformable successions within sedimentary strata. Due to the synchrony with the Triassic-Jurassic mass extinction, the CAMP volcanics have been studied extensively resulting in a number of studies that have better constrained the stratigraphic framework of the CAMP event (e.g. Blackburn et al., 2013; Dal Corso et al., 2014; Deenen et al., 2010; Nomade et al., 2007; Whiteside et al., 2007).

The volcanics of the CAMP have yielded a variety of ages across the four continents but all recent high precision U/Pb or $^{40}Ar/^{39}Ar$ ages are around 201±1 Ma (e.g. Marzoli et al., 1999; 2011; Nomade et al., 2007). Dykes and lava flows were dated at 199.8 ± 1.1 ($^{40}Ar/^{39}Ar$; Jourdan et al., 2009) in the Hartford Basin (USA), at ages ranging from 197.6 ± 1.1 to 199.5 ± 1.8 (Beutel et al., 2005; Hames et al., 2000) in South Carolina (USA), at 202±1 Ma (U/Pb; Hodych & Dunning, 1992), 202± 4 Ma ($^{40}Ar/^{39}Ar$; Dunn et al., 1998), 201± 2.5 Ma ($^{40}Ar/^{39}Ar$; Kontak & Archibald, 2003), 199.6 ± 0.6 ($^{40}Ar/^{39}Ar$; Jourdan et al., 2009), 201.38 ± 0.0.02 Ma (U-Pb; Schoene et al., 2010) in the Fundy Basin, at 199.9±0.5 Ma ($^{40}Ar/^{39}Ar$; Knight et al., 2004) and 199.1±1 Ma ($^{40}Ar/^{39}Ar$; Verati et al., 2007) in Morocco, and in Portugal at 198.1± 0.4 Ma ($^{40}Ar/^{39}Ar$; Verati et al., 2007).

The origin of the CAMP event, and its relationship with continental breakup, have long been debated. The estimated volume of CAMP material (>2.5 to 3 million km3) is immense and its distribution (>10 million km^2; McHone, 2003) is extremely wide. However due to post-depositional erosion of the CAMP volcanics, the volume of magma produced during this event is probably underestimated. In any case, such an event requires huge volumes of upwelling asthenosphere with very high rates of melting. Some authors believe that this event requires a plume head triggering regional doming with major thermal impact on the lithosphere (e.g. Hill, 1991; Ruiz-Martinez et al., 2012; Wilson 1997). Others argue that such large-scale magmatism is unlikely to be directly related to a plume, suggesting that the timing and geometric relationships do not support such an interpretation (Frizon de Lamotte, 2015; McHone, 2000) or that geochemical analyses suggest a magma source from the upper mantle with crustal contamination (e.g. Callegaro et al., 2013; Chabou et al., 2010). Some believe that it is closely related to oceanic spreading (Schlische et al., 2003) and may even have been the trigger mechanism (Courtillot et al., 1999).

2.4 Drifting and seafloor spreading

Seafloor spreading was previously thought to be synchronous along the entire eastern American margin and dated as late Early Jurassic to early Middle Jurassic (ca. 180-175 Ma: Klitgord & Schouten, 1986). More recently, drifting and the development of sea-floor spreading centres has been considered as being diachronous from south (Florida, South USA) to north (Orpheus Basin, Nova Scotia, Canada) along the margin (Klitgord et al., 1988; Schlische et al., 2003; Withjack et al., 1998) and occurring over a short time span. This interpretation is based on the observations that (1) inversion occurred just after the rift phase, during the Late Triassic in the Southern Segment and during the latest Early Jurassic in the Central Segment , (2) syn-rift sedimentation stopped and erosion occurred during early Late Triassic times in the Southern Segment and in mid Liassic times in the Central Segment, and (3) the Post-Rift Unconformity (PRU) is overlain by Early Jurassic sedimentary rocks in the Southern Segment and overlain by late Early Jurassic sedimentary rocks in the Central Segment. It appears therefore that drifting in the Southern Segment started in the latest Triassic to earliest Jurassic time (Schlische et al., 2003), and in the late Early Jurassic to early Middle Jurassic in the Central Segment (Schlische et al., 2003). However, other studies concerned with plate restoration prior to break-up do not agree on the diachronicity of sea floor spreading, and have dated the break up at 195 Ma in the Central Atlantic domain using magnetic anomalies and limits on salt deposition (Labais et al., 2010; Sahabi et al. 2004). After 195 Ma, ocean opening jumped to the east in the Tethys domain and sea floor spreading occurred in the Alpine Tethys region (Frizon de Lamotte et al., 2011). In the Northern Segment (between Newfoundland and Iberia), another phase of rifting started during the Late Jurassic (Sinclair et al., 1994), possibly commencing as early as the Early Jurassic (Pereira & Alves, 2013) with break-up in general occurring during the Early Cretaceous (Ziegler, 1988; Mauffret et al., 1989) and more precisely from 128 Ma (Barremian) between Grand Banks and the Portuguese margin (e.g. Peron-Pinvidic et al., 2007).

Withjack et al. (1995, 1998) showed that the prevailing tectonic regime changed during the transition from rifting to drifting in the Central Atlantic domain. In the Central Segment, the CAMP related dykes have a NE-SW trend. In the Southern Segment they trend either NW-SE or N-S (McHone, 2000). They cut pre-rift and syn-rift rocks and are not controlled by any pre-existing structures (Schlische et al., 2003). Based on dyke orientations, palaeo-stress has been estimated. The Early Jurassic stress in the Central Segment is consistent with a continued NW-SE extension. In contrast, the Southern Segment is incompatible with this and both sets of dykes cross-cut basin bounding faults at a high angle (Ragland et al., 1992). Accordingly Withjack et al. (1998) suggested that this was evidence for a significant change in tectonic regime and might represent the onset of inversion in the Southern Segment. They documented compressional structures in the Fundy Basin (**Fig. 2**) and in other North American basins where reverse faulting associated with inversion along rift-basin bounding faults followed normal faulting. Inversion in the Southern Segment is pre-CAMP and is post-CAMP in the Central and North Segments. Their interpretation of seismic data indicates that the northeast-striking boundary fault of the Fundy Basin experienced more than 4 km of reverse displacement after syn-rift deposition. A wide north-east anticline developed along the fault margin and the basin acquired a broad synclinal geometry.

3. Structural architecture

3.1 Structural compilation

Using published seismic interpretations and geological cross-sections from North America, Morocco and Portugal, we compiled a series of basin architectures in order to generate a large-scale view of Triassic basin structures and their extent (**Figs. 2, 3, 4**). In North America (**Fig. 2**), most Triassic basins have half-graben geometries. A few basins such as the Richmond and Orpheus Basins are full graben-like basins, and the basins in the northern segment (Newfoundland) are more complex and are difficult to image due to their depth and the presence of salt (Balkwill, & Legall, 1989; Schlische, 1993; Tankard et al., 1989; Welsink et al., 1989a, b; Withjack et al., 2010). Preserved basin dimensions vary greatly between 53 to 356 km in length and from 15 to 77 km in width. One basin (Pomperaug Basin) is significantly smaller with a length of 11 km and a width of 4 km (Schlische, 1993). However, it is important to recognise that these dimensions correspond to remnant basins. Triassic strata are tilted and truncated in the Southern and Central Segments (**Fig. 2**). The initial widths of the Triassic basins were much larger and many faults cutting the basin are post-Triassic. In the Northern Segment, most of the basin geometries are not preserved because of Triassic salt movement and diapir development during subsequent phases of margin evolution (**Fig. 2**), however in this area it is likely that the basins were wide and extensive.

Border faults are commonly segmented in the Southern and Central Segments. Fault segments overlap to form relay ramps, and fault strikes can vary (Schlische, 1993). The hanging wall margins of half-graben basins are commonly represented by small-scale, mostly antithetic faults, with minor throw (Schlische, 1993). In places, some basins display sedimentary wedges that thicken towards the border faults. However wedge development and basin architectures vary significantly between adjacent basins (**Fig. 2**). In the Southern basins, subtle growth-structures or areas lacking growth faults are present. The lack of evidence for significant syn-sedimentary faulting is important. Alluvial-fan deposits are often present in these basins which suggest significant topography during deposition. The Taylorsville Basin may be the only exception in the Southern Segment and shares more characteristics with the Central Segment basins.

In the Central Segment, growth-structures are more obvious (**Fig. 2**). In basins developed offshore eastern USA, wedges can be observed, but the basin-fills are not dated and the sediments could be Jurassic in age (Costain & Coruh, 1989). The very distal basins of the Scotian Shelf show small half graben geometries (10-25 km wide) and some of them display either an obvious syn-sedimentary wedge thickening towards bounding faults or more subtle growth in places (Welsink et al., 1989b). Truncation of the half-graben fills is widespread. An upper Triassic/ Hettangian unit was deposited above this regional surface and forms a tabular unit in very wide basins of regional-scale extent. Scotian shelf basins show a very similar architecture to the Moroccan basins (Hafid, 2000). Most of the Scotian offshore basins have the opposite polarity (basin-bounding faults dipping to the west), especially if they are in a distal position in the margin (e.g. Scotian Shelf basins). The very distal basins of the Scotian Shelf and the Orpheus Basin (in the Central Segment) are filled mainly with salt-bearing sediments such that the original depositional geometries are not distinguishable (Deptuck, 2010).

Along the Moroccan margin (**Fig. 3**), most of the sub-surface data is from the Essaouira Basin and to a lesser extent from the Souss Basin. Looking at E-W trending cross-sections, the basin architecture varies greatly from north to south. E-W trending profiles reveal half-graben geometries, mostly located in the east of the basins. In the northernmost part of the Essaouira Basin, Triassic deposits have tabular geometries and onlap basement to the north (**Figs. 3A, B**). Southward, Triassic deposits are extensive and local, small half-graben sub-basins (~ 10 km wide) are developed, displaying growth-structures along east-dipping faults (**Fig. 3C**). In the central part of the Essaouira Basin, many igneous sills are imaged and no wells have penetrated the entire sedimentary succession. Depending on which sub-basin is studied, 1, 2, 3 or 4 sills in the allegedly Triassic/ Hettangian sedimentary succession can be identified. Sills follow bedding in most places but may truncate bedding planes laterally. Attribution of a Triassic age to some of the lower sedimentary units is not always convincing. For example, in the Argana Valley, a thick Permian basin-fill succession has been described (Ait Chayeb et al., 1998; Medina, 1995) and Permian and Triassic deposits are conformable (see **Fig. 3E**). The sills imaged in seismic lines could either be related to the CAMP event intruding into the Permian succession or some of the interpreted lower igneous sills could in fact represent the Permian rhyolite lava. Consequently the lower sedimentary package beneath the sill could be Permian in age. In those basins without constraints from well data, sedimentary facies attribution is also not reliable. In the Central part of the Essaouira Basin, the architecture is complex. Some easterly-dipping faults were clearly active during deposition (**Fig. 3D**) but other faults have more complex geometries (**Figs. 3E, F**) with activity at the end of the Triassic (between coarser grained and finer grained facies as mapped from seismic data). However, some faults were active throughout deposition. In the very central part of the basin, a horst structure formed a structural high during deposition of a “coarse-grained facies” unit. The fault bounding the horst to the west seems to have been inactive during deposition of the “coarse-grained facies” unit while the faults to the east were active. The western fault appears to have been more active during deposition of the “finer grained facies” unit while the eastern faults were inactive. During the latest Triassic (Norian-Rhaetian), deposition occurred across a wide, extensive basin (> 60 km wide) and some authors refuted the existence of active fault during Triassic deposition (Baudon et al., 2012). In the Souss Basin, the western part of the basin was dominated by salt deposition directly on top of basement or Permian rocks. To the east, W-dipping faults formed small half-graben sub-basins with growth-structures along the faults. The basin fills (mapped as coarse-grained) were truncated creating a wide, extensive erosion surface on top of which fine-grained facies and salt were deposited.

In the Northern Segment, Canadian basins have thick salt packages and most of them have been displaced during diapir formation. Triassic sediments underneath the salt are not distinguishable and initial basin geometries not know (**Fig. 2**). However, from the salt distribution it can be inferred that the basins were very wide. In contrast, in the Portuguese basins (**Fig. 4**) a lower Triassic unit, the Silves Fm, is present below the Dagorda Fm, the evaporite-bearing unit, which is considered equivalent to the salt in the Canadian basins (e.g. Azeredo et al., 2003). In the Lusitanian Basin, the fine-grained Dagorda Fm does not contain much salt deposits, except offshore in the Peniche Basin, but it does contain dolomite (Alves et al., 2009). The lower Triassic unit (Silves Fm) is very tabular (**Figs. 4A, B**), thickening, locally in graben structures (< 10 km wide, **Fig. 4A**). To the SE of the Basin, half-graben geometries are present locally along a major basement fault zone (**Fig. 4D**). In these half-graben the Silves Fm is significantly thicker and the upper Triassic/ Hettangian unit (Dagorda Fm), where preserved, truncates the Silves Fm. Further off shore to the west and to the South, in the Peniche and Alentejo Basins, some half-graben basins with growth-structures are imaged seismically and show major displacement on some major faults, i.e Pereira de Sousa Fault and Sines Fault but most of the Silves and Dagorda Fm are tabular across the areas where major basins are formed during later rifting phases (Alves et al., 2009; Pereira and Alves, 2011, 2013).

3.2 Architectural and structural zones

The studied basins can be divided into four main groups with two additional sub-types depending on their stratigraphy and architecture, which vary systematically from south to north (**Fig. 5**). Type A basins are wide to very wide remnant basins (30 to 70 kms) that show growth-structures along a main border fault (such as Connecticut Valley basins - Pomperaug, Hartford and Deerfield Basins, offshore Georges Bank Basin (GBB),and the Fundy Basin). Sub-type AA basins, such as Newark Basin and offshore basins, show more subtle growth-structures than other type A basins. Type B basins are remnant basins that are narrow to medium size (10 to 25 km), and show either no obvious growth structures or very subtle growth-structures (such as Dan River & Danville Basins, Farmville, Scotsville Basin, Culpeper Basin, Gettysburg Basin and offshore Norfolk Basin). Sub-type BB basins, such as Taylorville, Richmond and Deep River Basins, show better growth-structures than other type B basins. Type C basins are relatively small half-grabens (5 to 20 km wide) with growth-structures overlain by very wide basins, often filled by salt-rich deposits (Scotian shelf basins and Moroccan basins). Some of the small half grabens are truncated beneath the salt succession but in other places they appear conformable. Type D basins are very wide remnant basins (> 50 km) displaying mainly tabular geometries developed beneath the salt deposits. Beneath the salt, grabens with slightly thicker deposits occur locally, and half-graben geometries which wedge toward faults are recognized but not frequent. Type D basins are located in the Northern Segment and correspond to the offshore Newfoundland basins together with the conjugate Portuguese basins where geometries are better defined (**Fig. 5**).

The basin types are spatially organized and structural zones are defined: The Type A zone forms part of the Central Segment, the Type B zone covers the Southern Segment and southern part of the Central Segment, the Type C zone covers part of the Central Segment and the Moroccan margin, and the Type D zone includes the Newfoundland-Portugal margins (**Fig. 5**). The boundaries between the “structural” zones roughly correspond to ancient first-order crustal lineaments. The main obvious boundary in the North is the Gibraltar transform zone, laterally known as Newfoundland Fracture Zone. Additional main lineaments include the South Atlasic Fault Zone that can be traced across the Atlantic domain to an onshore American lineament (Withjack & Schlische, 2005; **Fig. 2**) which is known as the Brevard Bowen Creek Fault Zone that runs south of the Newark and Gettysburg basins. In the Appalachian domain there are a number of pre-Triassic basement lineaments that have probably been re-activated and played a role in the initiation and development of Triassic basins (Withjack et al., 1998) and a similar situation occurred in Morocco (Le Roy & Piqué, 2001; Piqué & Laville, 1995).

4. General sedimentary and stratigraphic characteristics

4.1. Sedimentary analysis

The onset of basin development and infill in the Central Atlantic domain is slightly different from one basin to another due to variations in inherited structures, timing, paleogeography, and paleoclimate, but the main depositional phase, from Carnian to Hettangian, is similar with similar depositional facies present across the entire Atlantic domain.

4.1.1 North American basins

In the North American Triassic basins, sedimentary successions contain similar facies, summarized for all basins by Smoot (1991). Some of the facies associations he recognised are coincident with formal stratigraphic units but most are not. Four main facies associations have been distinguished: (1) alluvial fan, (2) fluvial, (3) lacustrine, and (4) lake margin.

4.1.1.1. Fluvial and alluvial fan facies associations

The fluvial successions were briefly described by Smoot (1991) using information compiled from different basins of the Newark Supergroup (summary in **Table 1 and Fig. 6**). A number of studies have described the fluvial deposits in the NA basins (Hoffman & Gallagher, 1989; Lee & Froelich, 1989; Olsen et al., 1989; Parker et al, 1988; Smoot, 1985; Smoot 1991; Smoot & Robinson, 1988; Textoris et al., 1986). Only a few however, have described the architecture and facies distribution, and suggested sedimentary models for the basal fluvial unit (Hubert & Forlenza, 1988; Leleu et al., 2009; Leleu & Hartley, 2010; Leleu et al., 2010; Leleu & Hartley, *in press*). These latter studies have been undertaken exclusively in the Fundy Basin as it contains the best exposures of all the North American Seaboard basins. In the Newark Supergroup basins, the basal fluvial unit has always been described as a fining-upward succession from conglomeratic to sandy (Leleu & Hartley, 2010; Smoot, 1991). Smoot (1991) summarized the fluvial and alluvial fan deposits from the Fundy and US basins and identified four alluvial fan facies and seven fluvial facies. We have reviewed these facies in **Table 1** and refined its nomenclature. We propose four main fluvial facies associations and three different types of sedimentary fluvial systems (see **Fig. 10**; two fluvial facies associations correspond to a single fluvial system and represent downstream/ upstream equivalents).

Smoot (1991) noted that the description and mapping of the fluvial units in many North American basins was very similar such that descriptions from the fluvial facies in the well exposed Fundy basin could be applied to all NA basins. We describe the fluvial successions in the Fundy Basin below, where six facies F1 to F6 and five alluvial fan facies (AF1 to 5) are documented (**Table 1**).

Overlying the basin margin unconformity or along an intrabasinal horst, coarse debris flow deposits passing into stream flow deposits (AF1 to AF4) and local colluvium deposits (AF5) are locally preserved (Leleu et al., 2009; Leleu & Hartley, *in press*; Table 1).

A coarse fluvial unit (< 110 m in the Minas sub-basin of the Fundy Basin) overlies the alluvial fan facies and comprises stacked fining-upwards cycles (3 to 10 m in thickness) of conglomerates to medium-grained sandstones (Leleu et al., 2009). Claystones and siltstones are occasionally preserved and occur as infrequent intraclasts. This fluvial facies is similar to Smoot's facies F1 to F3 and named "coarse fluvial deposits" in Table 1. The cycles can be traced laterally for over 10 km and up to 23 km forming a broad, laterally extensive braid-plain (Leleu et al., 2009). The width of the fluvial system and the radial palaeoflow distribution suggests a depositional model of a distributive fluvial system with one or more apices developed at the top of the half-graben hanging-wall dip slope. Channels are highly mobile, braided and probably anastamosed (Hubert & Forlenza, 1988).

A sandy fluvial unit (> 350 m in the Minas sub-basin), the base of which (> 40 m, probably 130 m in thickness) comprises medium to very-coarse sandstones forming multilateral stacked channel bodies with local preservation of claystone beds overlies the coarse fluvial unit. The upper part of this unit (> 210 m in thickness) comprises fine- to very coarse- sandstones forming stacked channel body complexes intercalated with clay-rich fine sandstones and thin claystone units (Leleu et al., 2010; **Table 1**). The channel body complexes are 6 to 20 m in thickness with a minimum lateral extent of five hundred metres but likely to be a few kilometres. The extent of the fine-grained units is unknown, but it is possible that they are laterally eroded and that channel body complexes are connected. In this unit, sandstones are mostly well sorted, trough-cross stratified medium-grained sandstones but some beds are coarse-grained to granular in places. This facies is not described by Smoot and mayt not crop out in the US basins (unless it partly corresponds to his facies F4-F5). The facies is interpreted to represent the distal part of a distributive fluvial system with braided and possibly anastamosing channels e.g. Hubert & Forlenza (1988). This is the distal equivalent of the coarse fluvial deposits. Fine-grained facies F6 correspond to the floodplain intercalated between channel complexes.

The upper part of the fluvial system is a mixed fluvial unit (> 170 m in thickness) that contains alternating channelized fluvial deposits which evolve upwards from braided to more sinuous channels, and unconfined flow deposits considered to represent splays (tabular fine- to medium-grained sandstones), local aeolian dune bedforms (coarse-grained well sorted sandstones) with interdune deposits, and playa units (tabular claystones) (Leleu & Hartley, 2010). The unconfined flow deposits are described in Table 1 as part of the lake margin and therefore are labelled LM1 and LM3. Such that in Table 1, this mixed fluvial unit is not included, as Smoot (1991) did not describe a fluvial facies that varies significantly and corresponds to the transition between fluvial and lake environments. However, it could correspond to his descriptions of sinuous fluvial channel deposits which form a thin unit at the top of the Wolfville Fm in the Fundy Basin (e.g. F4 and F5; Table 1). His description of F4 and F5 could also correspond to unconfined flow deposits, although, unconfined flow deposits in the Wolfville Fm, interpreted as terminal splay deposits, can be integrated into the lake margin facies described by Smoot (1991) (facies LM 1 and LM3, see below; **Table 1**). This unit is up to 55 m in thickness in the Fundy Basin. The lateral extent of the playa units are not known but probably pinch out up-dip and thicken down-dip.

4.1.1.2. Lacustrine and lake margin facies associations

Lacustrine conditions developed in all the North American Atlantic margin basins but occurred diachronously and with a variety of facies. Lakes developed in hydrologically-closed basins (Olsen, 1988; Olsen & Schlische, 1990; Smoot, 1985), and depositional environments related to lacustrine phases always show (1) a systematic increase in grain size towards the basin margins, (2) palaeoflows away from the basin margins, (3) the provenance of coarse-grained sediments is local, (4) many show frequent drying phases with the formation of evaporites, and (5) cyclicity within fine-grained deposits in the centre of the basin (deeper water facies) with considerable lateral continuity (Smoot, 1985). The lacustrine and lake margin facies are summarized in **Table 1 and Figure 6** following Olsen (1990) and Smoot (1991).

Three main types of lacustrine environments are identified: a deep perennial lake facies (L1 and L2), a shallow ephemeral lake facies (L3, L4, L6), and a playa facies (L4, L5 and L6; **Fig. 10**). In the northern basins (i.e. Fundy, Deerfield, and Hartford Basins) deep perennial lakes never developed. Where perennial lakes formed, deep water conditions initially developed and then shallowed upwards. Therefore, in a standard stratigraphic succession the deep lake facies consistently abruptly overlie the basal fluvial succession. The lacustrine facies then display a progressive transition towards an upper shallow lake unit (Olsen, 1990). Although two main lacustrine types are reported on the stratigraphic charts (**Fig. 6**), three end-member depositional models are described from facies associations (Olsen, 1990). The three end members are the Richmond-type, the Newark-type and the Fundy-type. The Richmond-type display facies corresponding to the deepest lake environment with more constant anoxic environment (black finely laminated or massive claystone and siltstone and gray sandstones; L1 and L2) containing fossil fish, plant debris and logs. The Newark-type facies associations comprise lacustrine conditions varying from anoxic (L1, L2) to more oxic conditions and occasional drying phases (red siltstone and sandstone with polygonal cracks) containing fossil fish, reptile footprints, and fossil plants. The Newark-type is the most common and occurs in the Dan River, Culpeper, Gettysburg, Newark, Pomperaug, Hartford and Deerfield basins. The Richmond-type is less common and occurs only in the Richmond and Taylorsville basins which probably had the deepest water depths (Olsen, 1990) and in the Deep River, Danville and Farmville basins, with the presence of coal beds. The Fundy-type model contains facies mainly related to playa, lake margin and shallow lacustrine environments, and contains fossil fish in places (Ackermann, 1995; Hubert & Hyde, 1982; Olsen et al., 1989). It is typical of the northern basins which become increasingly evaporitic towards the east and north. A fourth-type should be considered for the Scotian shelf and Newfoundland basins where evaporites largely dominate the fine-grained deposits. As mentioned above, the origin of the evaporites is debated. The lower sequence could be continental and the upper sequence could be marine (Jansa et al., 1980; McAlpine, 1990) but no definite agreement has been reached as yet.

The thickness of the lacustrine units in the upper sequence of the Triassic basins varies greatly from 260 m to 4900 m (Olsen, 1997; **Fig. 6**). The thickest lacustrine successions are located in the Central Segment between the Culpeper Basin in the south and the Newark basin in the north. However, lacustrine units of significant thicknesses (1300 m to 3000 m) were deposited in the Southern Segment, but have been partly removed by post-depositional erosion.

Lake margin deposits (LM1 to LM4) that include sandstones and coarser grained bodies intercalated with finer grained lacustrine deposits have been described (Smoot, 1991; **Table 1**). Some fluvial channels that have been incised into fine grained lake deposits can be observed and are considered to have formed when lake level fell (LM1). Depending on the type of lake and its depth, lake margin deposits vary from distal splay deposits (LM2 and 3) to fan-delta deposits (LM4 and 5).

4.2.1 Moroccan basins

The main outcrops of the Triassic Atlantic basins of Morocco are located in the Argana Valley which forms the western part of the Haha sub-basin in the Essaouira Basin. A synthesis of the main outcrops is given in **Figure 7** and a compiled stratigraphy of the Triassic in Morocco is given in **Figure. 9**. The first studies of the Triassic sedimentology were carried out in the Argana Valley by Ambroggi (1963), Tixeront (1973) and Brown (1980), and defined the lithostratigraphy of the Argana Valley and Essaouira Basin. Three main formations and eight members were defined initially as Triassic: the Ikakern Fm sub-divided into two sub-members (T1 and T2), the Timezgadiwine Fm (T3, T4, T5) and the Bigoudine Fm (T6, T7, T8) with a basalt bed on top (**Figs. 7, 9**). However the conglomeratic Ikakern Fm (T1 and T2) has subsequently been re-attributed to the Permian (Jalil & Dutuit, 1996) confirming a previous study by Dutuit (1976). These conglomerates are restricted to a graben structure in the centre of the Argana Valley (Brown, 1980). The thickness of the Triassic succession in the Argana Valley and Essaouira Basin varies greatly. It is thinner in the north (< 2000 m) and thickens towards the south with up to 4000 m in the Argana Valley itself (Hafid, 2000). The facies of the Essaouira Basin are documented in **Table 2** and summarized below.

The contact between the Ikakern Fm and Timezgadiwine Fm is marked by an angular unconformity. The Timezgadiwine Fm contains a thin fluvial conglomeratic sheet at its base (Tanameurt or T3 member). The conglomerates gradually thin to the north (Brown, 1980; Jones, 1975).

The Aglelgal Sandstone Member (T4) records the extension of sedimentation across the entire Argana Valley. It is 800 m thick in the north and up to 1500 m thick in the south. It lies either conformably on T3 or unconformably on Palaeozoic rocks. The Aglelgal Sandstone Member comprises alternating coarse-grained cross-bedded sandstone units and thick intervals of mudstone intercalated with siltstone and fine-grained sandstones. It has been interpreted as recording playa and sheetflood deposition (Hofmann et al., 2000).

The Irohalene Mudstone Member (T5) varies in thickness from 200 m to 500 m. The basal part of T5 is dominated by mudstones with abundant mottling. The upper part of T5 contains sandstone beds which increase in thickness upwards. T5 deposits are attributed to meandering ephemeral streams (Tourani, unpublished data). The boundary between T4 and T5 is debated (Brown, 1980; Hofmann et al., 2000; Tixeront, 1973) but the facies variation is probably gradual.

The Tadrart Ouadou Sandstone Member (T6; 0-150 m in thickness) comprises three distinct channel-like bodies located at different places in the Argana Valley (Table 2). The nature of the contacts between T5 and T6, and T6 and T7 are debated for which unconformities are suspected. This unit contains conglomerates, sandstones showing parallel and low-angle cross-bedding and climbing ripples and mudstones (Mader & Redfern, 2011). A southern unit comprises thin halite beds interbedded with sandstone and siltstone beds (Hofmann et al., 2000).

The T7 Member (up to 200m in thickness) is a fining-upwards sequence of graded sandstone beds at the base evolving to interbedded rippled siltstone and mudstone beds. Hofmann et al. (2000) suggest that T7 and T8 are similar and comprise cyclically-arranged mud-rich facies and cannot be separated into two distinct lithofacies units and interpreted those deposits as part of a playa system.

The Hasseine Mudstone Member (T8; 300-1100 m) was described by Brown (1980) as containing mainly claystones, siltstones and very fine sandstones with minor amounts of chlorite, halite, gypsum and anhydrite. Mudstones and siltstones are extensive and tabular. It is now accepted that the T7/ T8 deposits were formed in shallow ephemeral lakes, saline mudflats with periodic fluvial and aeolian inputs (Hofmann et al., 2000).

In the Argana Valley, a basaltic unit conformably overlies the top of T8 (Manspeizer et al., 1978; Tourani et al., 2000) and can be traced to the north within the Haha and Essaouira sub-basins. Correlations from Hafid (2000) show that the Bigoudine Formation (T6, T7, T8) dominated by claystones and siltstones in the Argana Valley becomes evaporitic to the north of the Essaouira sub-basin, where T7 and T8 are dominated by salt deposits. The basalt forms a marker bed between the lower and upper evaporite units (Ettouhami, 1994; Hafid, 2000; Salvan, 1984). The lower evaporite unit comprises alternations of anhydrite and halite beds, locally interbedded with potassic salts, and contains one thick massive salt layer. The upper evaporite unit comprises alternations of anhydrite, halite and locally potassic salts, as well as red claystone beds. Anhydrite is particularly abundant in the uppermost part of the succession (Ettouhami, 1994). To the south, the upper evaporite unit probably corresponds to the playa unit termed T9 by Hofmann et al. (2000), and overlain by a T10 member composed of fluvial conglomerates. Both T9 and T10 are dated as Early Jurassic in age.

Palaeoflow measurements in the Timesgadiwine and Bigoudine Formations indicate that sediments were derived from the east (Brown, 1980) and give no evidence of a structural high to the west. Within-graben palaeocurrents were consistently towards the west while closer to palaeo-slopes, sediments were transported northwards in the Argana and Tizi-Maachou grabens. A north-eastern limit to the Argana Basin is suggested by south-westerly dispersal.

To the North, in the Doukkala Basin, the data are poor but what is available for the depositional sequence shows a lower conglomeratic and sandy unit overlain by a siltstone-dominated unit in turn overlain by a thick salt package (the lower evaporite unit; **Fig. 7**).

To the south, in the Souss and Tarfaya Basins, data are sparse (only a few unpublished wells), and the main data are based on seismic interpretation which has identified a tectono-sedimentary package between the basement and the basalt and/or a well defined Lower Jurassic reflector. The main uncertainty in this basin is the precise age of this tectono-sedimentary package, particularly as the first package overlying basement in the Argana Valley of the Essaouira Basin is Permian in age.

In the High Atlas and Moyen Atlas domains, the typical Triassic succession is well exposed south of Marrakech in the Ourika Basin where growth structures have been described (Baudon et al., 2009). It comprises a basal conglomerate (termed F3), a lower siltstone unit (F4, also termed Ramuntcho Siltstone), a thick sandstone unit (F5, known as the Oukaimeden Sandstones), and an upper siltstone unit (F6) overlain by basalt (**Fig. 7**). Published sedimentological data and interpretations of this succession are contradictory, except for F3 which has been accepted as an alluvial fan deposit. The Ramuntcho Siltstone (F4) has been interpreted as a playa deposit (Beauchamps, 1980) and shallow marine due to the presence of bivalves, echinoderms, bivalves and algal mats (Biron & Courtinat, 1982). El Arabi et al. (2006) interpreted the base of F4 as fluvial and the upper part as lacustrine due to the presence of palaeosols, desiccation cracks and fresh water bivalves. The Oukaimeden Sandstones have been interpreted as a fully continental fluvial unit (Lorenz, 1976; Mattis, 1977), as deltaic and tidally-influenced deposits (Beauchamp, 1988; Benaouiss et al., 1988, 1996), and recently as a thick fluvial unit with a tidal bed at the top (Fabuel-Perez et al., 2009). The Upper Siltstone F6 was divided into two units by El Arabi et al. (2006) including the lower F6 unit with aeolian dunes and brown playa claystones and the upper F6 unit comprising siltstones and evaporites which were both interpreted as playa deposits but were interpreted as being lagoonal in earlier work (Beauchamps, 1988) and more recently as tidally-influenced (Konotio et al., 2015). F6 is supposed to be equivalent to the T7 and T8 members of the Argana Valley. In the Moyen Atlas region, the sedimentary succession comprises basal conglomerates, a lower siltstone formation containing dark or green claystones and locally evaporites, a basaltic unit and an upper siltstone formation.

4.3.1 Portuguese basins

The main Triassic sections in Portugal are exposed onshore in the Lusitanian and Algarve Basins. Both have similar depositional sequences (Palain, 1976) with the best studied being the Lusitanian Basin where most of the previous and recent work has focused (**Figs. 8, 9**). The two main sequences are the Gres de Silves Fm and the Dagorda Fm (Azerêdo et al., 2003; Palain, 1976; Soares et al., 1985; Soares et al., 2012; Uphof, 2005). Within these Formations, Palain (1976) defined three main units termed A, B, and C, sub-divided into A1, A2, B1, B2, C1 and C2. A1 corresponds to a coarse clastic fluvial unit (dominated by sandstones and conglomerates). A2 forms a fining-upwards sequence comprising sandstones, siltstones and claystones. Siltstone and claystone beds are dominant and tabular and sandstones are locally found in lenses. On top of A2, halite pseudomorphs were found (Palain, 1976). A2 would therefore correspond to splay deposits and playa margin environments. B1 corresponds to another fluvial unit but contains more pebbles than A1. B1 includes more conglomeratic facies that display either cross-bedding or a more massive fabric. At the base of B1 two conglomeratic units can be traced regionally. The B1 sandstone beds are more organized and form large cross-beds. On top of B1, a dark claystone containing plant fragments and *in situ* roots is present. This claystone is overlain by a sandstone bed, itself overlain by beds of shelly limestones. The base of B2 is a thin unit dominated by siltstones and claystones containing numerous plant fragments and root structures. In the upper part of B2, the siltstone and claystones beds are intercalated with dolomite beds and lenses and beds of sandstones. The sandstone lenses form fining upwards sequences containing abundant fossil debris at the base (plants, gastropods, and bivalves). The fossils include 20 species and are considered to be Hettangian in age (Boehm, 1903; Choffat, 1880). C1 overlies B2 and the contact corresponds to a regional erosional surface (unconformity; Soares et al., 2012). C1 is a thin unit of continental strata, starting with poorly sorted and massive conglomeratic beds. Laterally the conglomerates are absent but beds of sandstone that fine-upwards are present (Palain, 1976). Between sandstone beds, lenses of coal are frequent. C2 is dominated by horizontal and tabular beds of siltstone and dolomite intercalated with thin, fine-grained sandstones beds. C2 displays disturbed bedding, interpreted to be due to the dissolution of evaporites and possibly the presence of sulphates (Palain, 1976). Plants, coal debris and shells are locally abundant on top of the dolomitic carbonate beds or dolomite clay beds.

Palain (1976) undertook the main sedimentological and palynological studies in this area and combined them with previous work on bivalves and gastropods (Choffat, 1903). For Palain (1976), A1, A2, B1 were considered part of the Silves Fm and the top of B1, B2, C1 and C2 were part of the Dagorda Fm. Recent studies have attempted to update the work of Palain and divided the lithostratigraphy using similar boundaries between formations but renamed A, B, and C into the Conraria Fm (A1, A2), Castelo Viegas Fm (B1), and Pereiros Fm for B2, C1 and C2 (Rocha et al., 1990; Soares et al., 1985). The latest synthesis on the Triassic of Portugal (Azerêdo et al., 2003) further revised the stratigraphy considering the Dagorda Fm as a lateral/ distal equivalent to most of the Castelo Viegas Fm and the Pereiros Fm. Azerêdo et al. (2003) attributed ages to the different lithostratigraphic units, mostly using the work of Palain (1976). The northern Triassic sections in Portugal have been studied more recently in term of basin fill sequences bounded by unconformities or disconformities at the base and top of B1 with a main intra-B1 unconformity (Soares et al. 2012). Unit A is considered as one megasequence (namely Conraria Fm) while unit B is subdivided in 2 megasequences (corresponding to Penela Fm and Castelo Viegas Fm, separated by the unconformity). The two lower megasequences represent complex changes in the alluvial systems through time while the upper megasequence records the evolution of the fluvial environment towards more littoral facies in the uppermost deposits. It is interpreted to follow an important tectonic episode of the Lusitanian Basin (Soares et al., 2012).

In southern Portugal, outcrops south of the Lusitanian Basin at Santiago de Cacem, display a depositional sequence with a lower fluvial unit (conglomerates and sandstones), a middle unit dominated by siltstones and intercalated with thin sandstones beds, and an upper unit dominated by dolomite beds. In the Algarve Basin, the red bed sequence is divided into 5 depositional units. The basal unit is termed AA and is discontinuous across the basin, thinning out to the west (Palain, 1976). It comprises lenses of conglomerates and sandstones overlain by mudstone and siltstone beds intercalated with dolomite beds. The AB1 unit is dominated by sandstones and contains intercalated conglomeratic beds in the east. The AB2 unit is dominated by red and green mudstones. Fine-grained sandstone and dolomite beds are locally present. The AB3 unit is thin (few meters thick) but extensive and represents a marker bed in the basin. AB3 is a dolomitic unit formed by the intercalation of dolomitic mudstones, shelly dolomite and massive dolomite (Palain, 1976). Overlying this marker bed another unit contains red and green mudstones and local gypsum. CAMP volcanic rocks are present in the Algarve Basin but Palain's (1976) section does not report it. However he did mentioned dykes and sills intruding in the upper part of the Algarve red beds.

4.2. Biostratigraphy

4.2.1. North American basins

Accurate and precise age control is critical to understanding the extent of facies diachroneity from one basin to another and to assess correlations between basins as the onset of Triassic sedimentation is very poorly constrained. The transition between the fluvial and the lacustrine units is better constrained but is not always precisely dated. Stratigraphy and correlation within the North American basins have been developed using freshwater floral and faunal remains. Palynology is the most robust technique available to develop extra-basinal correlations in continental strata but well preserved tetrapod bones and reptile footprints have also proved useful in North American Triassic basins (Olsen et al., 1982; Olsen & Sues, 1986). Palynofloral zones are well established worldwide and NA basin fills contain relatively diverse preserved palynomorphs which allow the general definition of the stratigraphy of most basins (Cornet, 1977; Cornet & Olsen, 1985; Cornet & Traverse, 1975;). In lacustrine deposits of NA basins, fossil fish are the most abundant and diverse faunal remains and when combined with palynology provide a robust stratigraphic framework and correlation scheme (Cornet et al., 1973; Cornet & Olsen, 1985). However, most studies do not specify in which facies and where in the stratigraphy samples have been taken from. Macro- and micro- fossils are generally referred to the lithostratigraphic units to which they belong without specific location data.

In the NA basins fossil fish were used to define the stratigraphy and to develop correlation schemes in combination with palynology and terapod remains (Olsen et al., 1982). Five bio-stratigraphic zones have been recognized but only two are strictly Triassic age. The fish zones are the *Dictyopyge* zone dated Middle Carnian by palynology, and the *Diplurus newarki* zone dated Middle to Late Carnian. The three youngest fish zones are Jurassic in age (i.e. Hettangian, Late-Hettangian to Early Sinemurian, and Sinemurian) and are characterized by species groups of the holostean *Semionotus*.

Many authors have worked on tetrapod remains in the Newark Supergroup basins including: Anderson & Anderson (1970), Anderson & Cruickshank (1978) for the Early and Middle Triassic, and Olsen & Galton (1977), Olsen et al. (1982), for the Late Triassic and Early Jurassic fauna. These works were reviewed and compiled by Olsen & Sues (1986) and Olsen et al. (1987). Most tetrapod remains of Triassic age were reptiles, amphibians and mammal-like reptiles. The oldest specimens are strictly Anisian and Ladinian (*Proterosuchidae* and *Trematosauridae*) and come from an isolated outcrop in the Fundy Basin. Other specimens are from Anisian to early Norian (*Kannemeyeriidae, Rauisuchidae, Trilophosauridae,* *Tanystropheidae*), and from Carnian to Norian (*Rhynchosauridae, Metoposauridae, Capitosauridae, Stagonolepididae, Phytosauridae, Kuehneosauridae*, and *Drepanosauridae*) or Ladinian to Norian (*Procolophonidae*). Some taxa lasted from the Carnian to Jurassic but others appeared in earliest Jurassic times allowing the definition of the Triassic/ Jurassic boundary.

Cornet (1977) and Cornet & Olsen (1985) proposed seven pollen and spore zones with four specifically of Triassic age. The oldest zone, termed the Chatham-Richmond-Taylorsville palynoflora contains 155 species. Diagnostic-age species give an age of Middle Carnian to late Middle Carnian. The New Oxford-Lockatong palynoflora lacks Middle Carnian species and is interpreted as Late Carnian. The Lower Passaic- Heidlersburg palynoflora appears to be Norian. It changes upwards gradationally into the Manassas-Upper Passaic palynoflora which is probably Rhaetian in age. The latter zone ends at a rapid transition from a high diversity assemblage to a low diversity one where Triassic taxa are absent. This transition is located within an 80 m barren interval below the CAMP basalt in the Newark Basin (Cornet & Olsen, 1985).

A specific study in the Bay of Fundy suggested that the section in the Chignecto sub-basin contains Ladinian specimens at the base and Carnian specimens at the top (Nadon & Middleton, 1985). This section contains the oldest palynomorphs found in the NA basins.

4.2.2 Morocco

4.2.2.1. Essaouira and Doukkala Basins

Palaeontological finds constraining the age of the red-bed successions are relatively common in the Argana Valley within the Essaouira Basin. Initial studies of Triassic stratigraphy in the 1970’s used macro-fauna and were followed by palynological studies, and radiometric age dating of the volcanics.

The Ikakern Fm was previously considered to be Early to Middle Triassic in age based on vertebrate tracks and plant remains but a Permian age for the formation is now accepted (Jalil & Dutuit, 1996). The Timezgoudiwine and Bigoudine Formations are attributed to the Late Triassic using vertebrate remains and palynomorphs. Several studies dated the interbedded basaltic lava using radiometric methods. A recent study using Ar isotopes found ages ranging from 197.8 (± 0.7) and 201.7 (± 2.4) Ma with a peak at 199.1 (±1) Ma (Verati et al., 2007) concordant with U/ Pb age of 199.6 (± 0.3) Ma (Pálfy et al., 2000). To the north, in the Rif domain, marine limestones overlying the clastic rocks and basalt were dated biostratigraphically as Carixian/ Domerian (194 Ma, see Medina, 1995).

In the Argana Valley, Crustacea Phyllopodia specimens were identified in lower-basaltic units (Bigoudine Fm) and dated as late Carnian-early Norian (Defretin & Fauvelet, 1951). *Voltzia Heterophylla* was recognized (Koning, 1957) to be of Triassic age but no precise dates were given. In the same area, Duffaud (1960) found reptiles and fish bones and other authors (Brown, 1980; Dutuit, 1976a, b, c, d; Tixeront, 1973) found similar remains and trace fossils. Only a recent study (Jalil, 1999) gives a precise age of Middle to early Late Carnian to the T5 member in the Argana Valley and a new study gives an early Scythian age to various footprints in sediments initially believed to belong to the latest Permian Ikakern Fm (Tourani et al., 2010). Only Tourani et al. (2000) give a Late Carnian to Norian age to the T7 member of the Argana Valley based on palynomorphs.

In the northern basins which were "open" toward Algeria, dating within Upper Triassic beds has been undertaken using lumachelle (Rhaetian in Meseta area; Gentil, 1905) or in carbonate beds in easternmost Morocco (*Avicula contorta*; Muchelkalk; Lucas, 1942). A Keuper age has been recognized near Meknes (Zeida mines) by Ellenberger & Schmitt (1976) using ichnofauna and small proto-mammal bones.

4.2.2.2. High Atlas and Moyen Atlas domains

In the High Atlas area, despite the presence of an assemblage of macro-fossils (bivalves, bones, echinoderms, bivalves and algal mats) in the F4 formation of the High Atlas of the Ourika basin (Biron & Courtinat, 1982), no precise ages were assigned. However, in a basin to the north, Schmitt (1976) found ichnofauna and a proto-mammal bone, which suggested a middle to upper Keuper age. In the eastern High Atlas, the succession overlying the basalt contains bivalves dated as Hettangian (Rakùs, 1986). In central Morocco, Triassic *Voltzia Heterophylla* was recognized by Termier (1948) in some successions but no precise ages were assigned.

In the High Atlas, the Ramuntcho Siltstones (F4) have only been dated recently with a well preserved and rich palynological assemblage giving a Middle Anisian age for the base of F4 (El Arabi et al., 2006). The Middle Anisian age for the base of F4 was unexpected and therefore F3 must be either early Anisian or even Scythian (El Arabi et al., 2006). The Oukaimeden Sandstones have been dated by collection and recognition of palynomorph assemblages. Cousminer & Manspeizer (1976) sampled the upper part of the succession while Biron & Courtinat (1982) sampled the base and they all concluded a Middle Carnian age. The Upper Siltstone Formation has been dated as Carnian to Norian (Tourani et al., 2000) and the overlying basalt is considered to be Norian to Hettangian with respect to the age of the overlying dolomites in which Sinemurian palynomorphs have been found (Courtinat & Algouti, 1985).

Many studies dated the basalts in different basins of the Moyen Atlas using palynomorph assemblages, sampling sediments just below, within and overlying the basalt units. In the northern area of the Moyen Atlas, the succession below the basalt has been dated as Carnian whereas sediments interstratified with the basalt were considered to be Rhaetian-Hettangian in age (Baudelot & Charrière, 1983; Baudelot et al., 1990). In the north-west of the Moyen Atlas, sediments within basalts have been dated as lower to middle Norian (Baudelot et al., 1986). A recent study (Lachkar et al., 2000) shows that in the northern Moyen Atlas, beds beneath the basaltic unit are upper Carnian. In the southern Moyen Atlas, beds beneath the basalts are Norian and beds overlying the basalts are Rhaetian to lower Hettangian. In the central Moyen Atlas, beds within basalts were dated as upper Carnian to lower Norian. Lachkar et al. (2000) noticed that the pollen assemblages are related to the peri-Tethyan domain but one species is typically from Gondwana (i.e. *Samaropollenites speciosus*).

4.2.3 Portugal

The macrofauna of the Lusitanian sections were studied mainly between 1881 and 1948 and re-assessed during Palain's (1976) study. Macro-plants were found in A2 and at the base of C but the minimal number of species and lack of important stratigraphic markers make the stratigraphic interpretation difficult. In upper B2 and C, fragments of *Clasthropteris meniscoides Bong.* were found which is normally associated with Keuper/ Rhaetian facies in Europe but in this location it was associated with 20 species of gastropods and bivalves dated as Hettangian (Boehm, 1903; Choffat, 1880). *Voltzia ribeiroi Teix.* was also found but ranged from A2 to C within the section. In the Algarve Basin, the AB3 unit and uppermost AB2 beds contain bivalves and gastropods that are attributed to the Lower Liassic (Palain, 1976).

In lower A2, numerous palynomorphs were found and interpreted to be Late Triassic, pre-Rhaetian (Palain, 1976). In upper A2, the palynomorph association is different but is considered to be the same age (Adloff et al., 1974). On top of B1, the dark claystone containing plant fragments and *in situ* roots contains palynomorphs that were interpreted as Hettangian (Palain, 1976), but plant remains attributed to the Rhaetian had been found in this bed. Therefore Palain (1976) decided to locate the Triassic/ Liassic boundary above this bed - a few meters above the B1/ B2 boundary. Two other palynomorph associations were recognized in the upper B2 and C and were all interpreted as Hettangian (Palain, 1976). Work in the 1970's used a different timescale when referring to the Carnian/ Norian/ Rhaetian/ Hettangian boundaries, so the age of the studied successions cannot be compared directly.

To summarize, the A1 age is not constrained but is probably Late Triassic as is A2. Therefore A1 and A2 (Conraria Fm) are considered to be Carnian to early Norian (Azerêdo et al., 2003), B1 (Castelo Viegas Fm) is attributed to the Norian to Rhaetian. The upper Castelo Viegas Fm is believed to be a lateral equivalent to the base of the Dagorda Fm (Azerêdo et al., 2003). B2, C1 and C2 (Pereiros Fm and upper Dagorda Fm) are attributed to the Hettangian. In the Lusitanian Basin CAMP volcanics are not recognised but ash beds are present.

In the Algarve Basin, the lower AB2 unit contains palynomorphs and AB1 contain *estherii* that are attributed to the Upper Triassic (Palain, 1976). Palain (1976) mentioned CAMP intrusives (sills and dykes) in the upper part of the Gres de Silves Formation, but in his view, the intrusives cannot be older than the Liassic (due to cross-cutting relationships and biostratigraphy). However in the light of the modern stratigraphic framework, most of the Hettangian fauna described by Palain (1976) and previous authors are probably Rhaetian in age.

4.3 Stratigraphic correlations

Analysis of the biostratigraphic dataset (see below) suggests that the most reliable dates for the onset of Triassic deposition in the Atlantic domain are Carnian in the Southern Segment, Scythin to Ladinian in the Central Segment and conjugate basins in Morocco, and between Carnian or earliest Norian in the Northern Segment. Similar basin-fill facies are present within the different lithostratigraphic packages of each of the studied basins and can be grouped into fluvial, coal, deep lake, shallow lake, playa lake or evaporite units in the stratigraphic chart (Fig.9). Different lithostratigraphic units have been grouped into formations as defined by previous authors in each basin (**Fig. 6**). Most basins contain an infill history that shows an initial fluvial unit overlain by a lacustrine unit, the nature of which varies depending on the water depth in each basin. In addition, some basins show a thin basal fluvio-lacustrine unit and therefore two fluvio-lacustrine sequences were stacked during Triassic basin development. At the top of the Triassic basin-fill, the CAMP volcanic rocks are occasionally preserved but in many places dykes and sills cross-cut the sediments and no extrusive volcanic rocks exist due to post-extrusion erosion. The CAMP units which mainly form intrusive features within the basin fill successions are present as far north as the Jeanne d'Arc Basin and Portugal and as far south as Brazil and Algeria.

Olsen (1997) presented a stratigraphic chart that identified two phases of Triassic basin fill development (a lower and upper fluvio-lacustrine sequence) separated by an unconformity in most North American basins (**Fig. 6**). Each sequence was considered to fine upwards and record a change from fluvial to lacustrine sedimentation. However, stratigraphic analyses presented by Smoot (1991) and previous authors (Cornet, 1977; Cornet & Olsen, 1985; Olsen et al., 1989) did not recognize a lower sequence or an unconformity in most North American basins with the exception of the Fundy Basin (Nadon & Middleton, 1984, 1985; Smoot, 1991). However, it should be noted that in the Newark and Connecticut Valley basins a lower unit of unknown age has been seismically imaged (Withjack, pers. comm.). In the Fundy Basin, the lower unit is a Ladinian to Carnian fluvial sequence and although the relationship with the upper main sequence is unknown it can be interpreted as an unconformity (Withjack pers. comm., who has observed this relationship on one seismic line).

The thicknesses and duration of the different facies within each of the studied basins varies greatly with basin location (**Figs. 6, 7**). For example the fluvial units at the base of the main depositional phase range from 900 m to 2300 m. Lacustrine successions can vary from 560 m to 4900 m in thickness (**Figs. 6, 7**). Fluvio-lacustrine sedimentation was not synchronous across basins.

In the northern Central and Northern Segments, evaporites become dominant within the fine-grained lacustrine deposits, possibly as early as the Norian, but the age of the onset of lacustrine sedimentation is not known precisely. The basins located in the distal offshore areas contain thick evaporitic deposits organized into two main sequences, but their exact ages are not known. In addition, the origin of the evaporites is debated as geochemical signatures from the upper evaporitic sequence suggest a marine origin (probably from earliest Hettangian upwards) but the lower (Triassic) salt may be continental (Jansa et al., 1980; McAlpine, 1990). However there is no apparent agreement as to what geochemical signatures can be used to identify marine from non-marine salt.

In most basins, deposition was continuous through the Triassic/ Jurassic boundary and beyond except for some of the Southern Segment basins, where Triassic sediments were truncated due to a late Triassic phase of uplift and erosion (Withjack & Schlische, 2005). This prominent unconformity may have removed many kilometres of syn-rift and post-rift section.

5. Correlations between basins and across continents

5.1 North America

Lacustrine deposits record fluctuations in climatic and palaeoenvironmental conditions. In Triassic red-bed lacustrine sequences, cyclicity driven by climatic changes has been proven in the NA basins (Van Houten, 1962; Olsen, 1986, 1990, 1997; Olsen et al., 1989; Olsen & Kent, 1996) and has been used for correlation (Olsen et al., 1996). In the Lockatong Formation of the Newark Basin, Van Houten (1962) recognized two types of cyclicity based on detrital and chemical cycles that form rhythmic sequences attributed to wetter and drier climatic phases indicating expansion and contraction of the lake system. They form the smallest scale cycles described by Olsen et al. (1996), and named Van Houten cycles; they are 4 m thick in average and are governed by 20 ky astronomical cycles (e.g. Olsen, 1986). Olsen (1986) showed that 3 other scales of cycles exist and recorded several scales of Milankovitch periodicity. The shorter modulating cycle contain 4-6 Van Houten cycles and Fourier analysis indicates that it corresponds to short eccentricity 100-ky cycles. The McLaughin cycle exhibits 4 short-modulating cycles and corresponds to the 413-ky cycle which is the best expressed in the stratigraphy. The long-modulating cycle is made up of 5 McLaughlin cycles and corresponds to a ~2-my cycle. Combining cyclic analysis with magnetostratigraphy, Olsen et al. (1996) showed that these cycles are correlatable at the basin scale. He suggested an astrostratigraphic correlation method based on his definition of the cycles (assuming they represent a specific duration) which was then tied to a chronostratigraphic framework using magnetostratigraphy and the absolute age of the CAMP basalt (previsously dated as Hettangian and now revised to be Rhaetian). Initially, based on the palynology and astrochronolgy, the Newark Norian-Rhaetian boundary was considered to occur within chron E18 at 208 Ma whereas the Carnian-Norian boundary occurs within chron E13 at 218 Ma (Kent & Olsen, 1999; Olsen et al., 1996). Since the pioneering works of Olsen et al. (1996) and Kent & Olsen (1999), the Upper Triassic stratigraphy has been modified (see below) and the more recent 2012 Global Triassic Scale dated the Norian-Rhaetian boundary at 209.5 Ma, the Carnian-Norian boundary at 228.4 Ma, and the Triassic-Jurassic boundary at 201.3 Ma (Ogg, 2012).

The recognition of cyclicity within lacustrine stratigraphic sections has provided a useful method of correlation within a single basin (e.g. the Newark Basin; Olsen et al., 1996; Van Houten, 1962) and, in order to use cyclicity for developing a correlatable stratigraphic scheme for extending it at extra-basinal scale, it needs to be combined with methods such as magnetostratigraphy (Olsen et al., 1996), biostratigraphy, radiometric age data from volcanic rocks, as well as geochemistry (Blackburn et al., 2013; Deenen et al., 2010, 2011; Schoene et al., 2010; Tanner, 2010; Whiteside et al., 2007). Many recent studies have used these sets of methods in order to correlate the uppermost Triassic units (and mass extinction beds) across continents and correlated marine and continental realms (Dal Corso et al., 2014; Deenen et al. 2010; Ruhl & Kurschner, 2011; Schoene et al., 2010; Whiteside et al., 2010).

Correlation charts for the various continental NA Triassic basins are based on fish, reptile, and amphibian remains, palynomorph stratigraphy, geochronology, magnetostratigraphy, and geochemistry (mainly $ð^{13}C$). The CAMP basaltic units, dated by radiometric methods occur directly beneath the Triassic/ Jurassic boundary and form the only secure time line across the basins; although it is important to remember that some of the basalts are intrusive and form sills and dykes. These intrusives were the feeder systems for the overlying lavas, and the feeder systems are the only evidence preserved in many basins, such as in Florida. However where the CAMP units form flows within sediments, correlation across basins have been successful (Blackburn et al., 2013; Dal Corso et al., 2014; Deenen et al. 2010).

5.2 Morocco

Morocco contains the highest resolution biostratigraphic dataset, and a reasonably well-constrained East-West correlation panel (from the Essaouira Basin to the High Atlas and eastern Morocco basins) is based on biostratigraphy combined with lithostratigraphy (e.g. El Arabi et al., 2006; **Fig. 7**). A North-South correlation has been suggested by Hafid (2000) within the Essouira Basin but uncertainties remain due to the presence of various CAMP sills cross-cutting the bedding and others occurring concordantly within the stratified succession. These interbedded igneous rocks have been interpreted as extrusive volcanic deposits but examination of seismic lines indicates that they can be clearly identified as sills. In addition, seismic interpretations show lower Triassic units with significant thickening towards faults. It is possible that some of these units are in fact Permian in age, as recent work in the Argana valley which took place after the seismic stratigraphic interpretation, has shown similar wedge-shaped units of Permian age, and recent work on Triassic deposits has tended to minimize the thickness variation of the Triassic units across Argana (Baudon et al., 2012).

5.3 A global magnetostatigraphical scale for the Triassic?

Muttoni et al (2004) have used magnetostratigraphy to try to tie a high-resolution biostratigraphic (conodont biozones) framework and chemostratigraphically constrained section from the Tethyan domain (in Sicily) to the Newark Basin stratigraphic framework. Statistical analysis of the magnetostratigraphy from the two sections studied suggests that a change in the age of some palynofloral zones may be appropriate; however there are some significant uncertainties with the proposed correlation. These authors suggested that two statistically plausible solutions were possible: option 1 positions the Norian-Rhaetian boundary within E23 and the Carnian-Norian boundary within E15/16; option 2 (preferred by the authors) place the Carnian-Norian boundary within E7 (228-227 Ma), which implies a long Norian period with a time span of 20 m.y. and a Rhaetian period of 6 m.y. duration (Muttoni et al., 2004). They suggested that the Norian-Rhaetian boundary would be within E17 (209 Ma). A revised magnetostratigraphic framework proposed by Hounslow & Muttoni (2010) placed the Norian-Rhaetian boundary at the base of E21 and variably correlated this with the GPTS UT21-UT25. The last Global Triassic Scale (Ogg, 2012) placed the Norian-Rhaetian boundary at 209.5 Ma, the Carnian-Norian boundary at 228.4 Ma, and the Triassic-Jurassic boundary at 201.3 Ma, but the precise age of Upper Triassic stratigraphy is still debated (Ogg et al., 2014).

Magnetostratigraphy provides an excellent correlation mechanism for Triassic basins given that well exposed, continuous sections of relatively fine grained sediment are present with well constrained radiometric and/or biostratigraphic dates at the top and at the base of the sections. In reality, this situation is rarely achieved, so several sections are aggregated with three assumptions: 1) a composite magnetostratigraphic polarity profile can be determined (e.g. Newark Basin for which the magnetostratigraphy has been constructed using 10 ‘overlapping’ cores), 2) each section can be correlated, and 3) the section has not been diagenetically overprinted. In addition, the Triassic Global Magnetic Polarity Time Scale accuracy is uncertain (Ogg et al., 2014). Without an independent calibration of the age of the section, correlation is subjective and simply becomes a question of pattern matching (e.g. Muttoni et al., 2004). If the conditions are suitable for a magnetostratigraphic analysis then correlation between basins is possible.

5.4 Across the Central Atlantic domain

Recent studies applied magnetostratigraphy to define the Triassic-Jurassic boundary in Morocco and the Fundy Basin correlating Rhaetian strata between the two areas and interpreting the CAMP event as being responsible for mass extinction at the end of the Triassic (Deenen, 2010; Deenen et al., 2010). Earlier works have attempted correlations between the Newark Supergroup basins and Moroccan basins by defining large-scale sedimentary packages termed tectonostratigraphic sequences (TSI to TSIV) that are considered to have been “evidently produced by tectonic processes” and represent syn-rift sequences (e.g. Olsen, 1997; Olsen et al., 2000). Their analysis is supported by seismic interpretations and geological cross-sections in which sedimentary wedges are considered to thicken towards major basin-bounding faults as well as the observation that conglomerates located close to the faults indicate that local source relief was maintained throughout basin development (Cornet & Olsen, 1990; Costain & Coruh, 1989; Schlische, 1993; Withjack, 1995). The authors suggest that each tectonostratigraphic sequence can be assigned an age, and probably resulted from significant changes in rates/orientation of extension although they are not always bounded by unconformities.

TSI is the initial syn-rift sequence of Permian age (Olsen et al., 2000), previously thought to be Anisian (Olsen, 1997); TSII is an “early” syn-rift succession of possibly Anisian to Carnian age; TSIII a middle syn-rift succession of possible Late Carnian, Norian, and Rhaetian age; TSIV is the late syn-rift sedimentary and volcanic succession of early Hettangian to possibly Pliensbachian age (but volcanics are now revised as Rhaetian). Because each of these sequences is based on successions with poor age-control and on unconformities which are not always observable, the confidence in the definition of these sequences is limited. As the age of TSI is significantly older than the rest of the basin-fill succession it should not be included within the Triassic rifting evolution. Moreover, in many basins (e.g. the North American and Portuguese basins) there is a significant hiatus between TSI and the deposition of the first Triassic sediments. TSII is supposed to show a significant wedge-shape and rests unconformably on TSI or pre-rift basement. TSIII corresponds to the bulk of the basin fill in North America and Morocco (Olsen, 1997) and is considered to thicken towards the bounding faults. In the Fundy Basin TSIII corresponds to the fine-grained Blomidon Formation dominated by lake/ playa deposits (Olsen et al., 2000). It was in this interval in the Newark Basin where Olsen (1986; 1996) developed his astrostratigraphic correlation scheme. In Morocco, TSIII is represented by the Bigoudine Formation which comprises similar fine grained sediments to the Fundy Basin and overlies an unconformity (Olsen, 1997). However the basinwide unconformity in the Essaouira Basin at the base of TSIII is debated (Baudon et al., 2012; Brown, 1980; Hofmann et al., 2000). No evidence for large-scale fault activity has been recorded from the Moroccan field sections, only small-scale syn-sedimentary faults are reported (Baudon et al. 2012; Hofmann et al., 2000) which have been attributed to evaporite dissolution rather than tectonically-driven process (Hofmann et al., 2000). These authors suggested that the strata were deposited on a crustal domain where sedimentation was equal to subsidence. However Olsen (1997; Fig. 13) described an unconformity at the base of the Bigoudine Fm (TSIII) and seismic images show packages that could correspond to the TSIII unit that are conformable in places and unconformable in others (Hafid, 2000; see description above), despite these observations, the age of the unconformity remains unclear.

To summarize, the definition of TS as sedimentary packages should be used with caution particularly as the nature of the bounding surfaces varies greatly within one basin (e.g. sedimentary hiatus, disconformity, unconformity). In addition, changes in depositional style and lateral variation within a single basin coupled with age constraint uncertainties means that recognition of bounding surfaces can be problematic. When considering the Central and North Atlantic basins, Leleu & Hartley (2010) have observed that the sedimentological evolution of the basin-fill is similar in all basins but most likely diachronous. Therefore a fluvial dominated package in Morocco is not necessarily synchronous with a fluvial package in the Minas sub-basin, although in this case, the bulk of the fine-grained Blomidon Formation might have been deposited contemporaneously with the bulk of the fine-grained Bigoudine Formation in Morocco (**Fig. 9**). In addition it seems unlikely that fault activity in two different basins would produce exactly the same synchronous stratigraphic relationships.

5.5 Basalt units

Based on early radiometric data, Manspeizer (1981) and Salvan (1984) initially defined three Triassic and one Lower Jurassic basaltic event in Morocco. The first basalt is limited to eastern Morocco and was believed to be upper Ladinian in age. The second basalt is found across Morocco and was thought to be Carnian whereas the third basalt upper Norian. More recent work shows that there is in fact one Triassic basalt event which is dated as uppermost Triassic (e.g. Aït Chayeb et al., 1998). The latest one is Sinemurian in age but is limited to the High Atlas. A later eruptive phase is dated at 196.6 ± 0.6 Ma in the High Atlas of Morroco and at 194 Ma in the Fundy Basin (Jourdan et al., 2009) and may correspond to the last phase defined by Manspeizer (1981) and Salvan (1984).

Other studies comparing sections in the High Atlas of Morocco and the Fundy Basin show that the CAMP basalt occurred just after the Triassic/ Jurassic palynological turnover event and therefore was concordant with a very Early Liassic age (e.g. Whiteside et al., 2007). However, palynological data from the Moyen Atlas and eastern Morocco show sediments dated as Norian within some volcanic units. It is possible that the age initially determined from the palynological assemblages might be misinterpreted; reworked or localized volcanism occurred in north-east Morocco during the Carnian which if correct, would correspond to the first phase of volcanism identified by Manspeizer (1981) and would not be equivalent to the CAMP event.

6. Depositional environments

Triassic age depositional environments in the Central and North Atlantic domain consist of local alluvial fans, three different fluvial system types, and four lake types that are all of continental affinity although the salt lake type may be related to marine ingressions in the latest part of the Triassic (**Fig. 10**).

6.1 Sedimentology

6.1.1 Fluvial depositional systems

Studies of modern analogue fluvial systems from rift basins indicate that major fluvial systems do not flow into rift basins; in fact very few large rivers flow directly into rift basins (i.e. Rhine River) but more frequently flow away from the rift shoulders (e.g. Nile River and others in the E. African rift system, Lena River in Siberia). Fluvial systems that are developed in rift basins are either axial or more often transverse systems. The latter are represented by alluvial fans in small narrow basins, or in broader basins fluvial networks develop a distributive pattern where the river exits out the drainage basin (located on the hangingwall dip slope or rift shoulder (e.g. Gawthorpe & Leeder 2000; Leeder & Gawthorpe 1987) and becomes unconfined resulting in a decrease in sediment transport capacity, deposition and nodal avulsion. This is typified by the example of the Rio Grande rift Plio-Pleistocene fluvial system (Mack & Leeder, 1999).

The modern analogue datasets together with the Triassic examples from the Fundy (Leleu et al., 2009; Leleu et al., 2010), and Lusitanian Basins (Pena dos Reis & Pimentel, 2010), suggest that in most basins Triassic sedimentation was dominated by distributive fluvial systems (**Fig. 10A**). These fluvial systems are laterally extensive such as the example of the Wolfville Formation of the Minas Basin (Leleu et al., 2009) but also the Chinle Formation in the USA (Dubiel, 1992) or the Buntsandstein Formation in France (Bourquin et al., 2006). Our example of the Wolfville Formation shows a continuous transect through fluvial strata for over 45 km (Leleu et al., 2009). Because of the width and the radial palaeoflow of the

Minas Basin fluvial system, these distributive fluvial networks are the most likely to represent an analogue for the fluvial systems of the Triassic basins of the Central Atlantic domain (**Fig.10A**).

In the Triassic basins of the Atlantic domain, fluvial systems were perennial and dominated most of the basin area with main drainage systems being transverse to the basin margin and most likely joined a longitudinal system that could flow down the basin axis connecting a few sub-basins (**Fig. 10A**). However, the fluvial systems were never connected to an oceanic domain and were restricted to a few sub-basins. Other fluvial systems were smaller in size and ended in either terminal splays and shallow ephemeral lakes (**Fig. 10B**) or as lacustrine deltas in deeper lakes (**Figs. 10C, D**).

6.1.2 Lacustrine depositional systems

Four main types of lakes are distinguished: shallow perennial lakes (**Fig. 10C**), deep perennial lakes or anoxic perennial lakes (**Fig. 10D**), shallow ephemeral playa lakes (**Fig. 10E**) and salt lakes (**Fig. 10F**). Olsen (1990) described the first three types and these have been summarized previously. The origin of the salt lakes is more problematic insofar as their development could have been associated with a marine transgression. However, the palaeogeography suggests that any ingression would have had to have been extremely extensive as some of the salt basins are located a substantial distance from any coeval shoreline (up to 1000 km). In order to generate such a scenario, the floors of such basins would need to have an elevation close to zero or below sea-level and would be flooded frequently by sea water. Possible modern analogues include the Salton Sea in California and the Afar Depression in Ethiopia. However in both of these examples marine ingression would at most be only 100 km inland. An alternative way of producing thick packages of salt in continental basins is through continual groundwater recharge and evaporation as seen in many of the modern salt flats of northern Chile and Bolivia where up to a 1000 m of salt can accumulate in basins completely disconnected from marine influence. That is why we suggest two scenarios for Rhaetian palaeogeography (**Fig. 11, C1 and C2**).

The common lack of lateral control on sedimentary architecture and the necessary low-resolution bio-stratigraphic data available, mean that any understanding of facies variability is limited. Therefore the lithostratigraphic correlations on which most published schemes are based are bound to be inaccurate. Correlations between basins show that fluvial and lacustrine deposition occurs contemporaneously, but it is also necessary to understand the facies relationships within each basin in order to refine the palaeogeography. Triassic basins of the Central Atlantic domain are wide, numerous, and relatively close to each other. They are located far away from any marine basins, with the Tethyan Ocean being the closest marine system (at least a few 100 kms away). Therefore we can assume that it is most likely that the basins were not connected hydrologically, or that only immediately adjacent basins were connected (e.g. similar to modern Basin and Range basins).

6.2. Reconstruction of depositional palaeoenvironments

We have produced palaeogeographic maps (**Fig. 11**) based on plate reconstructions using the model of Sahabi et al. (2004) for the Central Atlantic (USA/ Nova Scotia and NW Africa) at 189.6 Ma. We tightened the plate reconstruction with estimated horizontal shortening for each basin but the reconstruction for the Newfoundland and Iberia margins is more difficult and less accurate. We produced palaeogeographic maps for the Ladinian (**Fig. 11A**), Mid Carnian (**Fig. 11B**), early Rhaetian (**Figs. 11C and 11CC**) and late Rhaetian (post CAMP event; **Fig. 11D**).

6.2.1 Fluvial distribution

6.2.1.1. North America

Conglomerates are mapped along the margins of the NA basins (details in Fundy Basin, Newark Basin, Deerfield and Hartford Basins, Culpeper Basin, Gettysburg Basin, Deep River Basin, Richmond Basin, Dan River Basin) mostly along the basin bounding faults. In some basins, conglomerates that represent alluvial fan deposits are also present along both margins of the basins (i.e. along border fault and uplifted hanging wall shoulder) (see Narrow Neck sub-basin in northern Gettysburg Basin, Culpeper Basin, and Deep River Basin, in Smoot, 1991; and Fundy Basin). In all basins, fluvial deposits onlap onto the hanging wall shoulder.

In the Newark Basin (and Narrow Neck sub-basin), Smoot (1991) suggests that fluvial systems both axial and transverse to the basin trough were present. Smoot (1991) reported that (1) fluvial deposits in the northern and southern end of the basin are laterally equivalent to lacustrine deposits and represent an axial drainage system, and (2) palaeocurrents indicating that the flow was from the southeast (across the basin and towards the border fault), indicates that a transverse drainage system was also present (**Fig. 11A**). He further suggested that the bounding fault was active (because of the presence of wedge-structure) but the fault did not block flow out of the basin. In the Hartford Basin, provenance and sparse palaeocurrent data suggest palaeoflow from the east (Hubert, 1978; **Fig. 11B**). In the Deep River Basin, palaeocurrent data indicate flow southward, parallel to the basin axis (e.g. Hoffman & Gallagher, 1989) and provenance data suggest source areas from the west or northwest initially, and later from the northeast or east (**Fig. 11B**).

6.2.1.2. Morocco

The fluvial successions in Morocco have not been studied in significant detail. The most extensive sedimentary studies took place in the Essaouira Basin (Brown, 1980; Mader & Redfern, 2011) and in the

High Atlas (Beauchamps, 1988; Benaouiss et al., 1996; Fabuel-Perez et al., 2009). In the Essaouira Basin, palaeoflow indicators show that the fluvial systems flowed from east to west with local deflection to the south in the Argana Basin for some of the stratigraphic intervals. In the High Atlas Basin, palaeoflow data suggest rivers drained to the east (**Fig. 11B**).

6.2.1.3. Portugal

All previous authors agree on the depositional environment in the Lusitanian Basin of Portugal. Alluvial fans and fluvial fans developed along the North-South trending topography east of the basin (e.g. Azerêdo et al., 2003; Palain, 1976; Pena dos Reis & Pimentel, 2010; Uphoff, 2005) and which formed the main topographic high separating the Lusitanian Basin from the Spanish Triassic basins. Uphoff (2005) suggested that the fluvial palaeoflow was constrained by structural features that focussed rivers between NE-SW trending horsts that formed sub-basins, and flowed to the southwest (**Fig. 11B**).

6.2.2 Lacustrine distribution

6.2.2.1. North America

Playa and deep lake facies are not developed synchronously within any of the Triassic basins, and no lateral transition from playa to distal anoxic shale transition has been recognised either. Deposition of these different facies occurred at different times but may occur in the same basin, with the shallow water facies passing up gradationally from the deeper water facies to form a drying upward sequence from the Carnian to Norian (Olsen, 1997). From the Norian to the Early Jurassic, facies suggest generally drier conditions (eventually with deposition of evaporites) in the north with deeper water lacustrine facies in the south (**Fig. 11C**). Anoxic conditions were developed in the southern basins (except South Georgia). The presence of coal beds only in the southern basins also attests to a climatic influence on sedimentation. The basins located in the east of the Central Segment (Deep River, Richmond, and Taylorsville Basins) and basins offshore and at the latitude of the Culpeper-Newark basins are likely to comprise deep lacustrine facies as well. In contrast, basins in the north are dominated by shallow water/playa lake facies, becoming increasingly evaporitic, with thick salt basins to the north and east from the Scotian Shelf northwards (**Fig. 11C**).

6.2.2.2. Morocco

In southern Morocco and the High Atlas, the first occurrence of lacustrine facies corresponds to mudflat, with some evidence of vegetation and evaporites. Evaporites became increasingly abundant through time. To the north of the Essaouira Basin, the evaporitic mudflat passed rapidly into a salt lake (**Fig. 11C**).

6.2.2.3. Portugal

The first lacustrine unit (A2; upper Conraría Fm), a lateral equivalent to the first fluvial unit (A1, lower Conraría Fm) in the Lusitanian Basin is a mudflat facies with no evaporates on this onshore part of Portugal (**Fig. 11B**) during Carnian times. The later lacustrine phase developed initially as a similar mudflat facies but became increasingly dolomitic through time (**Fig. 11C**). The presence of dolomite contrasts with the thick succession of salt developed offshore Portugal (Peniche Basin) and in adjacent basins in Morocco and Newfoundland at the same time. However, salt distribution is complex as there are salt deposits in the offshore Peniche Basins but not in the neighbouring southern Alentjo Basin.

6.2.3 Lacustrine facies and climate

The North American Seaboard basins contain a larger variation in the lacustrine facies than the Moroccan or Portuguese basins. This is partly due to climatic influence (evaporation versus precipitation) on the development of perennial versus ephemeral lakes. In the southern USA, the basins are considered to have been located within the tropical climate belt and more perennial lakes developed there (Olsen, 1997). Lake depth may have been controlled by structural architecture (Lambiase, 1990). To the north, lakes were mainly ephemeral and evaporites were formed at some point (**Fig. 11C**). The development of the later thick salt deposits may have been due to marine-related saline water incursion in lowland areas during deposition (**Figs. 11CC, 11D**). In contrast, salt precipitation in some of the earlier deposits may be linked to arid climatic conditions and the precipitation of salt through evaporative concentration of saline groundwaters (**Fig. 11C**).

7. DISCUSSION

7.1 Rifting processes

The mechanical behaviour of the lithosphere shows that a range of processes may be active during lithospheric stretching, i.e. uniform extension (e.g. McKenzie, 1978) or non-uniform depth-dependent extension (e.g. Royden & Keen, 1980), simple-shear (e.g. Wernicke et al., 1981) or detachment (Lister et al., 1986), or a combination of these processes, and consequently rift geometries can vary greatly. In addition, Buck (1991) defined rift modes, i.e. narrow, wide and core complex modes, based on rift geometries, dimensions and process duration, and he related each mode to mechanical characteristics of the lithosphere. Narrow rifting corresponds to stretching and necking of a strong plastic layer when strain is localised in the weakest zone; so narrow rifts are predicted to be prevalent when the crust is cold and the lithosphere is thick, typified by the Baikal Rift and the East African Rift systems. Wide rifts form due to specific and unusual lithospheric mechanical characteristics including high heat flow (hot crust), thin lithosphere (Buck, 1991; Buck et al., 1999), slow strain (Brun 1999) and a weak lower crust (Hopper and Buck, 1996; Huimans & Beaumont, 2014), which likely favour strain migration rather than localisation. When the lower crust is even weaker, lower crustal flow is efficient and high-grade metamorphic rocks are unroofed along major low-angle shear zones to form a core complex rift system (Buck, 1991; Dokka et al., 1986).

In the 90's, rift development of tectono-sedimentary sequences were modelled as a two-step process (e.g. Prosser, 1993) which included rift initiation passing into a rift climax stage with higher subsidence rates and followed by a two-fold post-rift process with decreasing subsidence (e.g. Gawthorpe et al., 1997; Gupta et al., 1998). The evolution from rift initiation to rift climax was interpreted to result from the linkage of small faults to form larger faults (e.g. Cartwright et al., 1995) generating increased subsidence, growth-structures and an overall deepening (lacustrine or marine) of sedimentary facies (e.g. Gupta et al., 1998; Watson et al., 1987). However, geometries in active rift basins can be complex when looking at sediments located near transfer zones or considering longitudinal sections (parallel to main faults) with depocenter shifts and disconformities between sedimentary packages (e.g. Morley, 1999). Many sedimentary based analyses of rift basins developed models based on a single basin architecture and grouped packages into syn-rift and post-rift sequences. However, recent studies tend to show that rifting might not be synchronous within pre-breakup margin rift basins (Cowie et al., 2015) or within a wide rift system (Stolfova & Shannon, 2009) and therefore the rift system has to be considered as a whole rather than as separate basins and/or intra-basinal sedimentary packages as is often done for narrow rifts.

7.2 Triassic basin development

The identification of the syn-rift and post-rift successions in North Atlantic Triassic basins has been debated over the years. Initially, coarse-grained units were thought to record syn-rift deposition whereas fine-grained units were thought to represent the post-rift (e.g. Whittaker, 1985), but later, it was demonstrated that the fine-grained units represented the main syn-rift phase in the UK and Irish basins because growth-structures along active faults were associated with the Mercia Mudstone Group and not with the Sherwood Sandstone Group (Lott et al., 1982; Jackson & Mulholland, 1993; Shelton, 1997; Ruffell & Shelton, 1999). Modern examples also confirm that fine-grained sedimentation (lacustrine phase) can occur during active phases of rift basin development (e.g. Lambiase, 1990); and most authors now agree that lacustrine deposits represent the main phase of rifting, such as in the Baikal rift basin, East African rift basins such as the Malawi and Tanganyika Lake basins (Hutchinson et al., 1992; Lambiase, 1990; Morley, 1988; Scholtz et al. 1998) and in China (Watson et al., 1987). Lambiase (1990) showed that fault activity and related topography are the main control on lake development due to segmentation of the rift basin. However in some basins, such as the Fundy Basin, growth structures are observed locally in different parts of the basin and associated with both coarse- and fine-grained units (e.g. Wade et al., 1996; Chignecto and Cobequid Faults), such that tectonic activity cannot be inferred from grain size variation, although in this basin the Triassic alluvial architecture clearly indicates syn-rift deposition and lacustrine deposits are considered to be late syn-rift (e.g. Leleu & Hartley, 2010). The overall upwards-fining profile seen in some Triassic basins can be explained by a decrease in source area relief through progressive erosion within a hydrologically closed basin. This, together with aggradation within the basin results in a decrease in the stream gradient profile, a reduction in the transport capacity of the fluvial system, an increase in suspended load and decrease in bedload (Smoot, 1991). In this scenario, catchment erosion outpaces source area rejuvenation, reflecting decreased tectonic activity. As the fluvial to lacustrine/playa facies transition is basin-wide, it effectively records a late phase of syn-rift sedimentation (Leleu & Hartley, 2010).

In North America, some studies show that some supposed basin-bounding faults were post-Triassic (Faill, 1973), as well as other fold and fault relationships (Hofmann et al., 2000; Lucas et al., 1988). Growth-structure interpretation and rift timing development have been criticized in North America (Faill, 1973; Lucas, 1988) and these studies have been overlooked in more recent work. Some of the concerns were about the excessive thickness of alluvial fan deposits which might mimic growth, whereas the bounding faults post-dated these fans (Faill, 1973). For instance, Faill (1973) believed that the Newark-Gettysburg Basin was not a fault-bounded graben but a "down-warp" due to continued crustal thinning during sedimentation. This description of basin evolution may be considered as the expression of a core-complex basin development with deposition of alluvial fan sediments occurring along a low-angle detachment fault. Lucas et al., (1988) suggested that most of the Newark Supergroup was deposited before the major Early Jurassic deformation. However, neither Faill (1973) or Lucas et al. (1988), were able to identify active faults. In the Fundy Basin, clear growth-structures are only locally observed from seismic data (i.e. Chignecto sub-basin and in the eastern part of the Fundy sub-basin). In the western part of the Fundy Basin no Triassic growth structures are observed. It should be noted however, that a fault bounds the deepest part of the Fundy Basin which could indicate that a major transverse fault might have been active during deposition and making the reconstruction of rift architecture difficult to unravel. In addition, post-Triassic fault movement in many basins makes it difficult to reconstruct original Triassic geometries.

The lack of growth structures in many of the NA basins does not necessarily indicate an absence of rifting; however the various geometries that do occur both spatially and temporally within the Triassic basins have to be explained. Variations in strain location can modulate the stratigraphic pattern within a basin-fill succession (e.g. Gawthorpe et al., 1997; Morley, 1999). It is well demonstrated for example that fault propagation can induce migration of the depocenter, as illustrated by both bed thickness variations and the formation of unconformities within basin fill successions (Morley 1999), making rift architecture complex. In graben-type basins (such as Richmond Basin, Southern Basins of NA), no obvious growth-structures would be expected anyway, but those Triassic basins are tilted following rifting and possibly prior to breakup, indicating that large-scale deformation can occur in the upper crust during the latest stage of rifting. More importantly, lithospheric stretching can happen through different processes, and it is likely that the mode of rifting (e.g. Narrow, Wide or Core Complex; Buck, 1991) could also explain the different basin geometries, and how they may vary from one mode to another through time and produce diverse stratigraphic patterns (e.g. Huismans & Beaumont, 2014).

In the Argana Valley (Morocco), some authors also suggested that growth-structures were misinterpreted (Baudon et al., 2012; Hofmann et al., 2000) and that unequivocal evidence for their recognition has not been observed (Manspeizer, 1988b). Those studies questioned the timing of rifting in Morocco as well as in North America. In Morocco, Hofmann et al. (2000) showed that variations in the thickness of T4 and T8 in the Argana Valley record compensatory shifts in depocenters during depositional cycles. Therefore local thickness variations do not affect the regional thickness of these deposits which show an overall constant thickness over tens of kilometres laterally and they suggested that T4 and T8 were deposited over a crustal domain that subsided with similar rates in both the northern and southern parts of the Argana Valley. This mainly tabular geometry with local subtle growth structures along faults is often seen on seismic data. In addition, Hofmann et al. (2000) suggested that some syn-sedimentary faults have affected the playa deposits of the Essaouira Basin and were formed due to evaporite dissolution. They refuted the idea that the Argana area was divided into horst and graben sub-basins (e.g. Brown, 1980). Hofmann et al (2000) suggested that two tectonic events affected the Argana Valley domain. The first deformation phase occurred prior to deposition of T3 and post-dated T2 deposition (which dates it as end Permian or Scythian). The second event post-dated T10 and pre-dated marine deposition (Hettangian; post-CAMP). Baudon et al. (2012) suggest that no major sedimentary wedges are recognisable from field data in the entire Triassic succession and that deposition occurred across a slowly subsiding area. However, due to the N-S orientation of the Argana Valley, Hoffmann et al. (2000) recognized that they could not exclude the possibility that growth-structures may occur westwards as suggested by Medina (1995). Present-day geometries of along-strike active rift basin like in east Africa show such sag geometries while dip-sections clearly show growth towards faults (e.g. Sander & Rosendhal, 1989). Hafid (2000) showed a seismic line from the Argana/ Essaouira Basin in which N-S profiles show little deformation but E-W or NW-SE profiles show unconformities and growth-structure within the Triassic succession (**Fig. 5**) attesting of tectonic activity, most likely between T3 and T6. However, at the base of the succession, some growth-structures are interpreted within narrow half graben, but some of these basal units are not convincingly Triassic in age and could be the Ikakern Fm, of Permian age. Good examples are given in the Permo-Triassic architecture of the NE Atlantic margin (Štolfová & Shannon, 2009) that illustrate a lower seismic unit (likely to be Permian) that wedges towards bounding faults and upper units showing shifts in depocentres and local subtle growth-structures or an entirely tabular geometry.

Nevertheless, in Morocco, some of the earliest Triassic deposits undoubtedly filled small half-graben and display growth-structures laterally to more tabular extensive units. A major event occurred after deposition of the lower unit, which resulted in tilting of the half graben fill and the development of a regional erosion surface and locally conformable on top of which seismic interpretation suggests packages of fine-grained Triassic strata and salt deposits are preserved. However, it is difficult to be certain of the interpretation of the seismic facies and the age of tilting and surface development. This surface could be Triassic (between T5 and T6?) or Hettangian (post-T10) as Hofmann et al. (2000) suggested. We interpret this surface as the much discussed unconformity present between T5 and T6. This event also occurs offshore Nova Scotia where similar architectures are preserved within small half-graben truncated beneath an extensive surface overlain by salt and deposited in much broader basins.

The geometry and structural development of most Triassic basins in Newfoundland and Portugal had been largely overlooked until relatively recently (Azerêdo et al., 2003; Soares et al., 2012; Uphoff, 2005), however it seems clear that the Silves and Dagorda Formation in Portugal have a rather tabular geometry when considered at a regional/ seismic-scale (Alves et al., 2003; Alves et al., 2009; Pereira & Alves, 2011; Pereira et al., 2013). However, localised wedge geometries and thickening towards basin-bounding faults have occasionally been observed beneath the Jurassic (Alves et al., 2003; Pereira & Alves, 2013). Similar to Morocco, observations of depocenter shifting in the onshore Lusitanian Basin have been recorded from the Silves Formation (Matos et al., 2010).

The Central Atlantic Triassic basin geometry described here is similar to that recognised by Štolfová & Shannon (2009) for Permo-Triassic basins of the NE Atlantic margin. They documented a large variability in basin architecture using recent seismic data. They showed 5 types of geometries: 1) basins with sedimentary wedges along faults, 2) tilted basins with no wedges (tilting post-Triassic), 3) very wide basins with an overall tabular shape, that also display small-scale wedges along different faults which maybe associated with depocenter shifts, 4) very wide, perfectly tabular basins overlying small-scale half-graben with growth-structures (suspected to be Permian), and 5) basins showing a subtle wedge and overlying smaller scale basins that fill a pre-existing paleo-topography (Fig. 9 of Štolfová & Shannon, 2009).

The sedimentary pattern of most Triassic basins of the Central and North Atlantic domain can be summarised in a simple way: a similar two-phase (fluvial to lacustrine) lithostratigraphic development. In some basins, an occasional lower fluvio-lacustrine sequence has been described and interpreted to be a first pulse of fault activity (Lambiase, 1990) in the rifting evolution. The main (upper) fluvio-lacustrine development was then deposited and comprises a lower part composed of coarse-grained fluvial deposits with rare occurrences of aeolian deposits, passing upwards into a finer grained permanent lacustrine or ephemeral playa unit. This change in sedimentary style occurs in all of the basins but the timing of the transition and the duration of fluvial and lacustrine sedimentation vary significantly between basins (**Fig. 9**). The fluvial-to- lacustrine transition occurred during the Anisian in the Irish Sea and English basins (e.g. Leleu & Hartley, 2010), in the Carnian in the Newark basins, in the Norian in Fundy and at the Norian/Rhaetian boundary in Portugal. Although exact dates are not known for the transition within most of the southern basins and those of offshore Ireland, it is likely that this transition is diachronous. For example, where age dates are more precisely constrained across England, the contact between the arenaceous Sherwood Sandstone Group sediments and the argillaceous Mercia Mudstone Group sediments is known to be diachronous (Ruffell & Shelton, 1999). The similar but diachronous nature of the stratigraphic evolution within the Triassic basins of the Atlantic margin suggests that global climate change (e.g. Ruffell & Shelton, 1999) was not the causal mechanism behind this gross facies transition.

To summarise, Triassic basins are considered to be generally very wide (> 70 km wide) and form large, broad depressions on the continental crust (**Fig. 12**), that extended for hundreds of kilometres across the centre of Pangea. They locally have bounding faults and growth-structure wedges, but most of the sediments were deposited in subsiding crustal basins with only limited fault control and partly formed by large-scale regional/ continental subsidence, especially during the late rifting stage. Locally, re-activation of major pre-existing structural weaknesses occurred in the upper crust and these were active during most of the rifting phase. Elsewhere, new faults formed to accommodate deformation as horst or half-graben structures over a limited time period. During a later syn-rift stage, most faults were probably not the main subsidence driver.

The Triassic Atlantic domain rift basins show a similar development to narrow rifts in the rift initiation stage with the reactivation of crustal lineaments (such as the Headlands & Chignecto Fault in the Fundy Basin or Nazaré Fault in Portugal) and the eventual development of upper crustal faults. The basins do however show a marked difference in the rift climax stage as many of the basins do not contain thick lacustrine successions reflecting deepening, so either the underlying mechanical behaviour does not allow strain localisation (necking) generally associated with higher subsidence or fault linkage did not happen. Whilst it should be noted that stratigraphic architectures that are related to fault linkage in large basins bounded by a major fault array (e.g. Morley 1999) do occur in some place (e.g. Argana and Fundy), these are not the predominant style in the Atlantic domain. The absence of fault linkage might be explained by a lack of coupling between the elastic-brittle lithosphere and an underlying high-viscosity layer which limits strain localisation (Heimpel and Olson 1996), and therefore fault linkage.

In most recent continental rift basins (e.g. Lake Baikal, Lake Malawi, Lake Tanganika) the lacustrine phase is considered to be the main syn-rift phase (Lambiase 1990; Morley, 1988). These examples are from narrow rift systems and therefore the sedimentary expression of such a system is likely to be different to wide rifts, where strain is distributed over a much wider area. In the wide rift zone of the Triassic Atlantic domain the lacustrine phase probably also represents a syn-rift phase but occurs during the latest period of the syn-rift phase (e.g. Leleu & Hartley, 2010). The long time-scale for Triassic rifting (> 35 My), the prevalence of continental deposits and basins which show a change from overfilled or balance filled (dominated by fluvial processes) to underfilled (dominated by lacustrine deposits) through time (e.g. Caroll & Bohacs, 1999) can be explained by the decrease in source area relief through progressive erosion within a hydrologically closed basin (e.g. Leleu & Hartley, 2010). It could also be argued that basins become underfilled after a phase of higher subsidence, resulting in spill points being located at higher altitudes with associated basin isolation. Whatever the overall cause in terms of the transition to lacustrine dominated sedimentation the model of deepening and basin linkage during the rift climax (e.g. Gawthorpe & Leeder, 2000) does not appear to apply to the Atlantic Triassic domain.

7.2. Continental-scale Triassic rifting model

Based on the oldest preserved syn-rift sequence, onset of deposition in Triassic basins varies from Late Scythian or Anisian in the Central Segment basins to Carnian in the Northern and Southern Segment (**Fig. 13**). The Central Segment is bounded by major structural transcurrent lineaments that can be recognized in North America and Morocco (see Withjack & Schlische, 2005 and **Fig 11**). We suggest that the Central Segment formed the initial nucleus for subsidence and deposition (**Fig. 13**). The major structural lineaments also seem to control basin type distribution (**Fig. 5B**) as the basins of the Central Segment are very different to the other neighbouring basins.

The Central Segment contains the first recorded Triassic deposition which was largely fluvial, and the initial strain of the Central Atlantic domain was accommodated through reactivation of major basement structures (**Fig. 14A**) as seen in the Fundy, Newark and Moroccan basins. For example, in the Fundy Basin, the bounding fault is a pre-existing thrust that was re-activated during the Triassic as indicated by the development of sedimentary wedges. In the Central Segment, offshore Nova Scotia and in Morocco, small half-grabens were created but age constrains are poor. Some half-graben basins are clearly truncated and depositional areas were initially larger and not bounded by the presently preserved fault (**Fig. 14B**). We suggest that basins that display half-graben geometries at present were not necessarily developed initially as half-graben, but have been subject to later faulting which has created such geometry. It is difficult to determine for how long faults were active, but from the limited data available it appears to vary significantly, with some faults active throughout deposition (inherited major lineaments) whereas others were active only at the start or the end of basin development. Following the phase of small half-graben development, a major change occurred that corresponds to an overall change in basin geometry and sedimentation and is associated with creation of a major truncation surface. Seismic interpretation suggests this change occurred roughly between deposition of the coarser (fluvial) and finer (lacustrine) grained facies. At that point basins became much wider, fault activity decreased and lacustrine facies dominated (**Fig. 14B**). However, in some fault-bounded basins, growth-structures are associated with the fine-grained deposits. Locally, truncation may occur above part of the half-graben fill of the earlier Triassic unit, but laterally deposits are conformable with the half graben fill. This phase of basin development occurs either before or during the deposition of the first thick evaporite unit (possibly mid Norian, late Norian or early Rhaetian). During finer grained facies deposition, depocentres shifted across some basins and sub-basins, suggesting that faults might have been activated locally, whilst others became inactive; such a scenario was described in the North Atlantic domain by Štolfová & Shannon (2009).

In the Northern Segment, rifting started during Carnian or Norian times. Pre-existing faults in Portugal may have also formed the locus of deformation with associated thicker sedimentary succession in these areas. During this time, deposition occurred across the entire Central Atlantic domain, basins were wide and tabular overall. Faults were active very locally and formed associated small half-graben. It appears that most Triassic sedimentation occurred within large broad basins where faults may have only acted to locally partition areas of the basin. In the Southern Segment, preservation of initial basin geometry is poor and it is difficult to constrain basin development from the truncated remnant basins in this area. Consequently the main focus of discussion on basin development concerns the Northern and Central Segments.

Overall, the continental-scale architecture shows two end-member basin types: 1) Half-grabens with obvious growth-structures along bounding faults; 2) Late Triassic wide rift geometries typified by large-scale tabular units, local depocentre changes and subtle thickening toward faults. Wide rift evolution is considered to be driven by lower crustal flow within a high heat-flow region (Buck, 1991; Hopper & Buck, 1996), and could account for the development of regional subsidence without hinterland rejuvenation for the fluvial drainage in the Central Atlantic domain. Considering the development of half-graben basins, two processes are distinguished: 1) Re-activation during extension of long-lived major basement lineaments, and 2) Adjustment of the upper crust relative to a maximum extension in the lower crust by small, local half-graben which correspond to brittle deformation of the upper crust accommodating a small amount of extension. During the late Norian and early Rhaetian, most of the basins experienced lacustrine (or lagoonal/paralic) sedimentation (**Fig. 9**). The sediments of this latter syn-rift phase overlie a major basinal unconformity, with topographic highs gradually overlapped and eventually buried (**Figs. 14B and C**). In order to be able to bury old topography and flood the topographically lowest basins with marine waters, uplift must have decreased (little fault activity) and subsidence must have been ongoing although subsidence rates may have decreased. This indicates that regional subsidence was important relative to sedimentation (e.g. Doglioni et al.1998). Therefore during late syn-rift basin development, thermal subsidence is considered a first order control, as opposed to tectonic subsidence.

The major basinal unconformity recognized across the wide rifted zone could represent an uplift phase during this long-lasting rift period as modelled by Huismans & Beaumont (2014). In their dynamic model of rifting they managed to reproduce architectures similar to the Triassic rift architecture comprising faulted early syn-rift basins and late syn-rift, extremely wide sag basin with little deformation, and filled by fluvio-lacustrine or shallow marine sediments. In the model of Huisman and Beaumont (2014) continental mantle or lower crust was replaced by hot asthenosphere beneath the large rifted area with limited magmatism during rifting. They show that there is no uniform extension within crust and mantle as in the McKenzie (1978) model but the lower crust or lithospheric mantle are being removed during rifting and before breakup. They mention that a hiatus in subsidence or uplift could occur when lithospheric mantle is thinned and replaced by asthenosphere.

Triassic rifting of the Central Atlantic occurred over a period of >35 My and evolved eventually into oceanic sea floor spreading at 195 Ma (Sahabi et al., 2004). Although rift-related volcanics have not been reported a massive magmatic event occurred at the end of the rifting (~202-198 Ma) leading to the CAMP lava flows and sills which many have speculated caused onset of continental breakup in this domain (e.g. Courtillot et al., 1999). The timing of the magmatic event following a long-lived rifting period suggests that the Triassic rifting is of passive type initiated by lithospheric extensional stresses causing mantle uplift that eventually lead to magmatic production from different sources (Beutel et al., 2005; Frizon de Lamotte et al., 2015; Nomade et al., 2007; Şengör & Burke, 1978; Ziegler & Cloething, 2004; **Fig. 14C**). Geochemical data from CAMP basaltic dykes of the southeastern American margin argue in favor of an upper mantle source for the magma (Callegaro et al., 2013) as well as data from Brazil (Merle et al., 2011), and Algeria (Chabou et al., 2010). In the southern basins, erosional truncation indicates a Late Triassic pre-CAMP uplift phase as well as on the southwest Iberian margin (Inverno et al., 1993). We suggest that this is related to the onset of mantle uplift that initiated surface uplift. Further to the north, a Late Triassic unconformity within the Nova Scotian and Moroccan basins might be related to large-scale pre- CAMP uplift of lesser amplitude. We suggest that the large-scale Triassic rift architecture is controlled by thermal effects related to the rift mechanism. The erosion event may correspond to lower crust or lithospheric mantle removal and asthenospheric replacement described in the model of Huismans and Beaumont (2014). Classically, long-lived rifting processes eventually triggered asthenospheric doming recorded by the topography, subsequent rift architecture and sediment infill associated with high heat-flows ultimately generating mantle melt to produce the CAMP (Frizon de Lamotte et al., 2015). Many authors suggested that the Pangea supercontinent had a warming effect at the base of the lithosphere as an explanation for the high melting rate and high magmatic production during the short CAMP event (Anderson, 1982; Beutel et al., 2005; Coltice et al., 2007; Gurnis, 1988; Zhong & Gurnis, 1993). At that time lava flows were interbedded with continental deposits only a few million years prior to sea floor spreading; the lithosphere was still buoyant despite being already thinned by rifting processes. This implies that from the end of the Triassic to the onset of sea floor spreading (only few millions years) rapid subsidence from roughly sea level to -2500m occurred after extreme thinning of the lithosphere (Cowie et al., 2015). Studies of various passive margins and their pre-breakup tectonics (Bache et al., 2010; Chabouraud et al., 2013; Cowie et al., 2015; Moulin et al., 2005; Stica et al., 2014;) show similar extreme subsidence after long subaerial deposition during rifting, although subaerial salt deposition could occur in basin at 300m below sea-level (Cowie et al., 2015). However most models fail to explain the processes that could trigger lithosphere buoyancy during rapid thinning of the lithosphere before the extreme subsidence (Lavier & Manatchal 2006), but new models give some solutions (e.g. Cowie et al., 2015; Huismans & Beaumont, 2014).

8. Conclusions

We present a compilation and synthesis of published structural, sedimentologic, and stratigraphic data in order to investigate the tectono-stratigraphic evolution of the Triassic basins in the Central and Southern North Atlantic domain. Published cross-sections and seismic reflection data show that only a few basins had long-lived growth structures during Triassic basin development. These long-lived bounding faults were pre-existing crustal weaknesses that localized and focussed initial extensional deformation. The geometry of the basin-fill in most basins is typified by large-scale tabular sedimentary sequences with subtle thickening of sequences toward faults. Basins are truncated indicating that the initial basin was significantly wider than the preserved remnant observed today. Four main structural types of Triassic basins were recognized and distributed in specific areas of the Atlantic domain: 1) the Southern domain characterised by small to medium size basins with limited growth-structures; 2) the Central-East domain comprises wide basins with growth-structures associated with long-lived bounding faults; 3) the Central-West domain includes a wide early basin-fill unit with numerous active half-graben sub-basins and a late very wide tabular basin-fill unit; 4) the Northern domain with very tabular, extremely wide basins. Overall, the architecture suggests that subsidence has to be of a regional extent and not controlled only by faults with deep crustal inheritance as first-order control on subsidence and therefore basin development.

We suggest that subsidence was driven by lower crustal flow within a high heat-flow region and that variations in crustal characteristics could lead to the development of the four main basin types.

We reviewed sedimentologic and stratigraphic data and compiled a series of correlation charts to conclude that basins were all filled by continental deposits in various palaeo-environments such as alluvial-fan, mega-fluvial systems, terminal fluvial systems, fan-delta fluvial systems, permanent oxic and anoxic lakes, playa lakes, and salt lakes, some of them punctuated by aeolian deposits. Palaeogeographic maps were generated in order to summarize conceptual palaeo-environment and facies distribution at a continental-scale. A typical gross depositional sequence shows a lower fluvial unit overlain by an upper lacustrine unit. Basins contain either one or two depositional sequences. Based on the best correlations that we can currently infer from various datasets, the onset of deposition and extension occurred significantly earlier (Scythian to Anisian) in the Central Segment (Newark to Nova Scotia – Morocco) than in the southern and northern segments where it occurred during the Carnian. This also underlines the role of structural inheritance on Triassic basin development. Two major transcurrent lineaments are identified: the Gibraltar Fracture Zone and the South Atlasic fracture zone which both have a role at continental scale and segment the basin architecture and basin evolution.

Climate also had an impact on Triassic sedimentation. Indeed, lake types are not only dependent on the basin structures controlling water depth in lakes but they are also distributed depending on climate zones. Perennial lakes dominated the southern domain whereas playa lakes and associated evaporites developed mainly to the north. Salt deposits are thick in some northern basins, located offshore Canada and in the more northern Moroccan basins, but uncertainties concerning evaporite origin remain; salt precipitation could be attributed to a climatic origin (arid, hydraulically closed basin) or a marine ingression. In the Canadian offshore basins, two main periods of salt precipitation are recognized but salt distribution varies greatly towards Portugal. Indeed, there are significant thicknesses to the southern basins offshore Portugal (Peniche and Algarve Basins) while onshore Portugal, there is no thick salt but interbedded evaporates, dolomites, and claystones. These are associated with marine fauna in the latest Rhaetian/ Hettangian times, which suggests that the second phase of salt precipitation is related to marine ingression, although the origin of the first phase (late Norian? Early Rhaetian?) is not constrained. In addition, it is interesting to note that when the Lusitanian Basin became fully marine, fauna is dominantly of boreal affinity, and only a few Tethyan species are present (Soares et al., 1993). Thus, although the initial ingression is Tethyan, the Tethyan influence decreased in favour of flooding from the north.

Previous work has suggested that the fluvial and lacustrine successions record changes in climate. Some authors support that lake development suggests a wetter climate but we consider that this assumption cannot be supported. The sedimentary evolution from a fluvial phase to lacustrine phase occurred diachronously in all the studied continental basins. In addition, basins were probably endorheic (hydrologically closed) for most of the time. Accordingly the alternation of fluvial/ lacustrine deposition more likely reflects large-scale basin development (rejuvenation/ uplift of rift shoulders, subsidence, sediment and water supply, fluvial gradient) rather than a climatic change. However climate will determine which type of lakes developed (perennial versus ephemeral, and possibly saline).

Tectono-stratigraphic analysis of the Central and North Atlantic basins shows that the Triassic rifting is a process that occurs at a continental scale. Preserved Triassic basins are very wide and it is likely that they were significantly wider than we observe at present. Some basins have tabular geometries at a continental scale, wedges along faults are subtle and fault activity was short-lived. The timing of basin development was diachronous and no single time period of rift initiation can be established nor the distinct pulses of rifting present in one basin are correlatable with those in adjacent basins because subsidence in many basins is not controlled solely by fault activity. Therefore strain was not necessarily accommodated by basin-bounding faults such that subsidence was largely driven by crustal processes. Because the Triassic rift system is controlled by very wide rift mode, the McKenzie model (1978) is not applicable and the "syn-rift" basin-fill geometries are complex.

Long-lived Triassic rifting in the Central Atlantic domain is an example of passive rifting with a late magmatic event leading to emplacement of the CAMP, one of the largest volume LIPs on Earth. We believe the onset of the CAMP event is represented by the occurrence of a major erosional truncation surface in the southern domain and an erosional unconformity (or locally disconformity) in the northern domain in continental deposits beneath the basaltic volcanic rocks. While uplift and erosional processes occurred in the south, large scale subsidence took place in the north with the accumulation of lacustrine and evaporitic deposits. The rifting phase evolved into oceanic spreading during Early Jurassic times immediately after the CAMP event. Nevertheless processes governing subsidence are not well constrained as large-scale continental rifting occurred on a thick lithosphere followed by localised, rapid subsidence in some basins immediately prior to the onset of sea-floor spreading.

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List of figures

Figure 1: Location of Triassic basins. (A) Plate reconstruction and palaeogeography of the Triassic rift system in the Central Atlantic domain (from Ziegler, 1990, modified); (B) Location of the studied Triassic basins on the plate reconstruction at 186.9 Ma, based on the model of Sahabi et al. (2004). The numbers (1 to 17) in the basins refer to stratigraphic columns (Figure 11). GNFZ: Gibraltar Newfoundland Fracture Zone; SAF: South Atlasic Fault.

Figure 2: Structural overview of the American Seaboard Basins from published cross-sections; (i) localisation map; (A) Jeanne d'Arc Basin (Tankard et al., 1989, modified); (B) Whale and Horseshoe Basin (Balkwill, & Legall, 1989, modified); (C) Orpheus Basin to Abenaki Basin; (D) Orpheus Basin to Sable Basin; (E) Emerald Basin to Mohican Basin; (F) Mohawk Basin (B to F from Welsink et al.,

1989a); (G) Chignecto to Fundy Basin (from Withjack et al., 1998, modified); (H) southern Chignecto to Fundy Basin (from Schlische 1993, modified); (J) central Fundy Basin (Withjack et al., 2010, modified); (K) Nantucket Basin to Atlantis Basin; (L) Nantucket Basin; (M) Atlantis Basin; (N) Long Island Basin; (O) New York Bight Basin (from Costain & Coruh, 1989); (P) southeastern Newark Basin (from Schlische, 1993, modified); (Q) central Newark Basin (from Schlische, 1993, modified); (R) Gettysburg Basin (from Schlische, 1993, modified); (S) Culpeper Basin (From Schlische, 1993, modified) ; (T) Taylorsville Basin; (U) Deep River Basin (from Schlische, 1993, modified); (V) Richmond Basin (from Schlische, 1993, modified).

Figure 3: Structural overview of Morocco from published cross-section in the northern Essaouira Basin known as the Abda sub-basin (A), (B) (C) (from Hafid, 2000, modified); in the Essaouira sub-basin (D) (E), (F), (G) (from Hafid, 2000, modified); in the Souss Basin (H), (J) (from Hafid, 2006, modified); (i) localisation map .

Figure 4: Structural overview of Portugal from published cross-sections; from the Lusitanian Basin (A), (B), (C), (D), (E) from Rasmussen et al. (1998), modified; (i) localisation map.

Figure 5: Definition of a structural scheme for the Triassic basins of the Central Atlantic domain (A) Four types of basins and two sub-types are defined based on structural architecture: Type A: wide basins with growth structures along long-lived faults; Type AA: medium to wide basins with subtle growth structures along long-lived faults; Type B: Medium basin with no obvious growth structures, Triassic basins are truncated; Type BB: Medium basins with subtle growth structure, Triassic basins are truncated; Type C: wide and rather tabular basin with underlying small, local half grabens displaying growth structure; Type D: extremely wide tabular basin with no growth structures; (B) Distribution and structural zonation of Triassic basins. Basin locations shown on plate reconstruction of Sahabi et al. (2004), modified. BBCFZ Brevard Bowen Creek Fault Zone ; GNFZ: Gibraltar Newfoundland Fracture Zone; SAF: South Atlasic Fault; circled numbers 1 and 2 correspond to basins in Gulf of Mexico and South Georgia respectively, the stratigraphy of which is given in Figure 9.

Figure 6: Composite sedimentary sections of the North American Seaboard Basins. Synthetic sections are based on description of Smoot (1991) and Olsen (1997); bed relationships and facies thicknesses are schematic, only formation thicknesses are accurate. Correlation are based on biostratigraphy and geochronology (Cornet; Smoot, 1991; Olsen, 1997; Kent & Olsen, 1999; Jourdan et al., 2009; Deenen et al., 2010 and see text for details) and the associated numbers indicate the level of confidence in the

correlation (5 is high, 1 is low). Correlation based on lithofacies associations cannot be done from published work.

Figure 7: Composite sedimentary sections of the Moroccan basins. Synthetic sections are based on sedimentary descriptions in Cousminer & Manspeizer (1986), Benaouissa et al. (1996), Hafid (2000), Lachkar et al. (2000), El Arabi et al. (2006). Correlations are based on biostratigraphy (Jalil, 1999; Lachkar et al, 2000; Tourani et al., 2000; El Arabi et al., 2006; Tourani et al., 2010 and see text for details) & geochronology (Vérati et al., 2007) and the associated numbers indicate the level of confidence in the correlation (5 is high, 1 is low).

Figure 8: Composite sedimentary sections for Portugal in the Lusitanian Basin, based on Palain (1976) and & geochronology based on Vérati et al. (2007).

Figure 9: Lithostratigraphic compilation of the 16 studied basins using GTS 2012 for time scale. Unconformities annotated by a star are based on Olsen (1997).

Figure 10: Schematic 3D perspective views of sedimentary models for the different environments found in the studied Triassic basins of the Central Atlantic domain. Model A: fluvial-dominated basin; Model B: basin dominated by mudflat and splay deposits; Model C: Shallow oxic lake; Model D: anoxic lake basin; Model E: basin dominated by Playa lake; Model F: Salt lake basin. The cartoons emphasize the distribution of various facies associations and their relationships. The fluvial system types: F1 Mega fluvial system, F2 Terminal fluvial system, F3 Fan-delta fluvial system. The various types of lakes: L1 Permanent lake, anoxic bottom, L2 Permanent oxic lake, L3 Playa lake (mudflat and occasional evaporites), L4 Salt lake.

Figure 11: Palaeogeographic maps (A) Ladinian; (B) Carnian; (C1) Late Norian to Rhaetian pre-CAMP with the hypothesis 1 that salt comes from groundwater evaporation where basins are far away from any marine influence ; (C2) Late Norian to Rhaetian pre-CAMP based on the hypothesis that salt is generated through evaporation of sea water following marine ingressions into lowland basins; (D) Rhaetian post-CAMP.

Figure 12: Schematic representation of large-scale architecture of Triassic basins across (A) the Central Atlantic domain (Central Segment), (B) the southern North Atlantic domain (Northern Segment) during the Rhaetian.

Figure 13: Map showing lateral variation in timing of the onset of the Central Atlantic Triassic rift and inherited structural controls.

Figure 14: Schematic upper crustal cross section model of tectonic evolution. Large-scale rifting evolution of the Central Atlantic domain during the Triassic. A) Phase 1: Activation of pre-existing crustal weaknesses. B) Phase 2: wide rift mode; 2a) strain is distributed in the crust and when the lower crust reaches a threshold, the upper crust adjusts and strain is accommodated by brittle deformation 2b; 2c) More distributed extension occurred again and lacustrine/ salt deposits accumulated. Shifting in depocentre localization and no obvious fault activity. C) Phase 3: Central Atlantic Magmatic (CAMP) event at 201 Ma followed by extra-thinning of the lithosphere eventually leading to sea floor spreading in the Early or Middle Jurasssic. However, in the southern North Atlantic, a more complex rifting event occurred and sea floor spreading only started during the Berriasian.

Figure 1

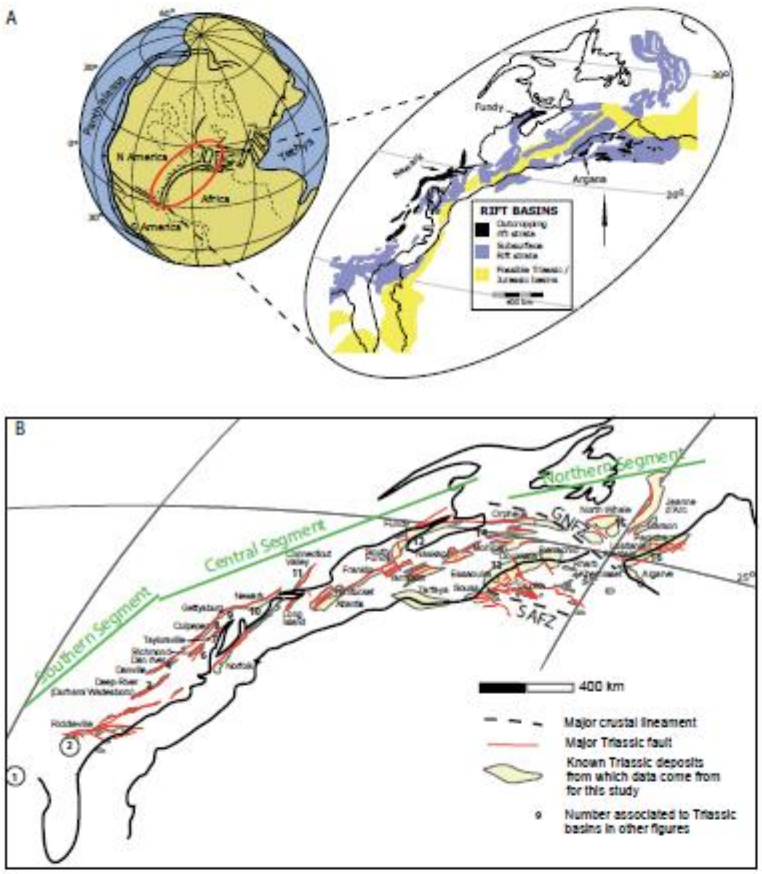

A
N America
Africa
S America
Tethys
Fundy
Newark
Argana
RIFT BASINS
B
Northern Segment
Central Segment
Southern Segment
GNFZ
SAFZ
400 km
Major crustal lineament
Major Triassic fault
Known Triassic deposits from which data come from for this study
Number associated to Triassic basins in other figures

Figure 2A

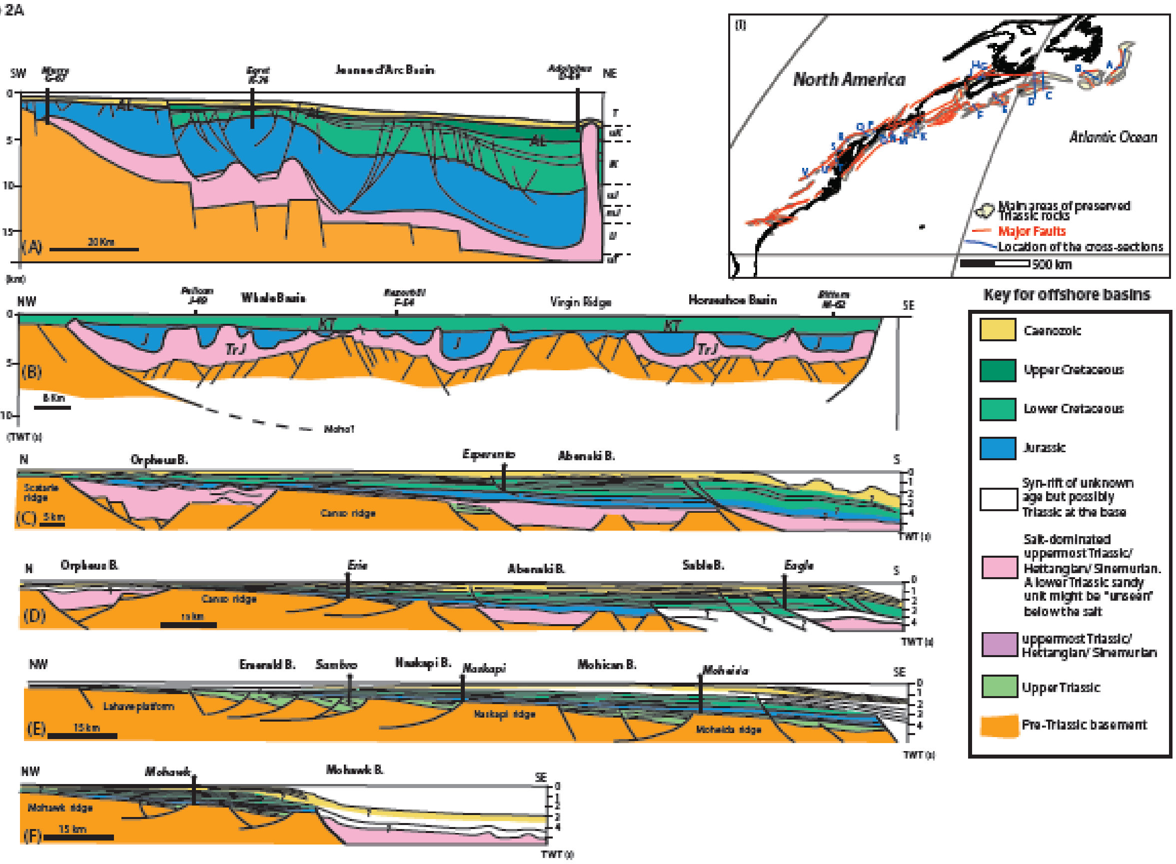

North America
Atlantic Ocean
Main areas of preserved Triassic rocks
Major Faults
Location of the cross-sections
500 km
Key for offshore basins
Caenozoic
Upper Cretaceous
Lower Cretaceous
Jurassic
Syn-rift of unknown age but possibly Triassic at the base
Salt-dominated uppermost Triassic/ Hettangian/ Sinemurian. A lower Triassic sandy unit might be "unseen" below the salt
uppermost Triassic/ Hettangian/ Sinemurian
Upper Triassic
Pre-Triassic basement
Jeanne d'Arc Basin
Whale Basin
Virgin Ridge
Horseshoe Basin
Orpheus B.
Abenaki B.
Canso ridge
Sable B.
Lahave platform
Naskapi ridge
Mohican B.
Mohawk B.
Mohawk ridge

Figure 2B

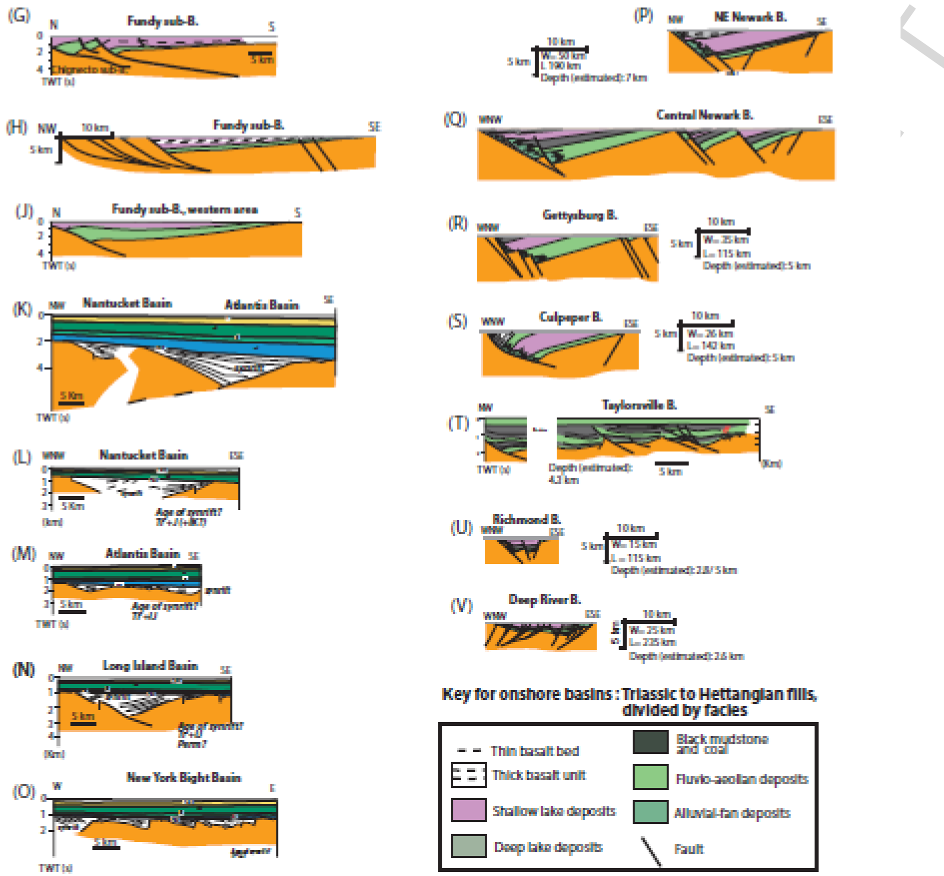

(G) Fundy sub-B.
(H) Fundy sub-B.
(J) Fundy sub-B., western area
(K) Nantucket Basin Atlantis Basin
(L) Nantucket Basin
(M) Atlantis Basin
(N) Long Island Basin
(O) New York Bight Basin
(P) NE Newark B.
(Q) Central Newark B.
(R) Gettysburg B.
(S) Culpeper B.
(T) Taylorsville B.
(U) Richmond B.
(V) Deep River B.
Key for onshore basins : Triassic to Hettangian fills, divided by facies
Thin basalt bed
Thick basalt unit
Shallow lake deposits
Deep lake deposits
Black mudstone and coal
Fluvio-aeolian deposits
Alluvial-fan deposits
Fault

Figure 3

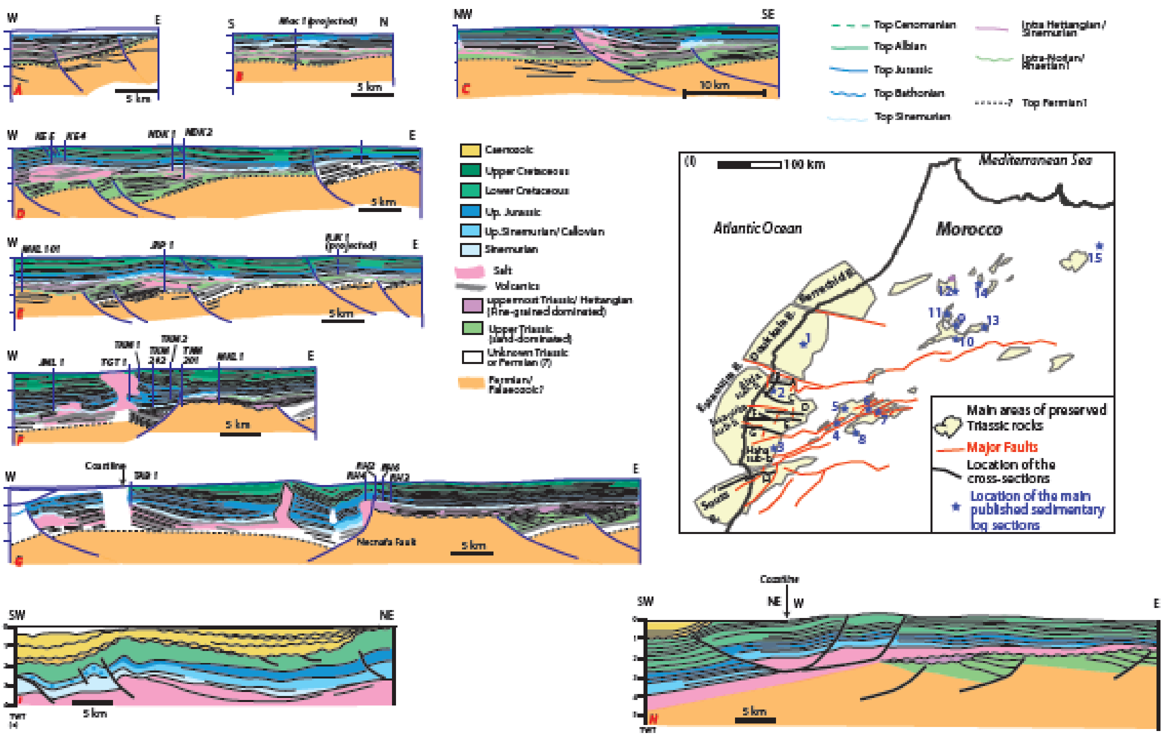

W
E
S
N
NW
SE
SW
NE
5 km
10 km
Top Cenomanian
Top Albian
Top Jurassic
Top Bathonian
Top Sinemurian
Intra Hettangian / Sinemurian
Top Permian?
Caenozoic
Upper Cretaceous
Lower Cretaceous
Up. Jurassic
Up.Sinemurian/Callovian
Sinemurian
Salt
Volcanics
uppermost Triassic/ Hettangian (fine-grained dominated)
Upper Triassic (sand-dominated)
Unknown Triassic or Permian (?)
Permian/ Palaeozoic?
100 km
Mediterranean Sea
Atlantic Ocean
Morocco
Main areas of preserved Triassic rocks
Major Faults
Location of the cross-sections
Location of the main published sedimentary log sections
Coastline

Figure 4

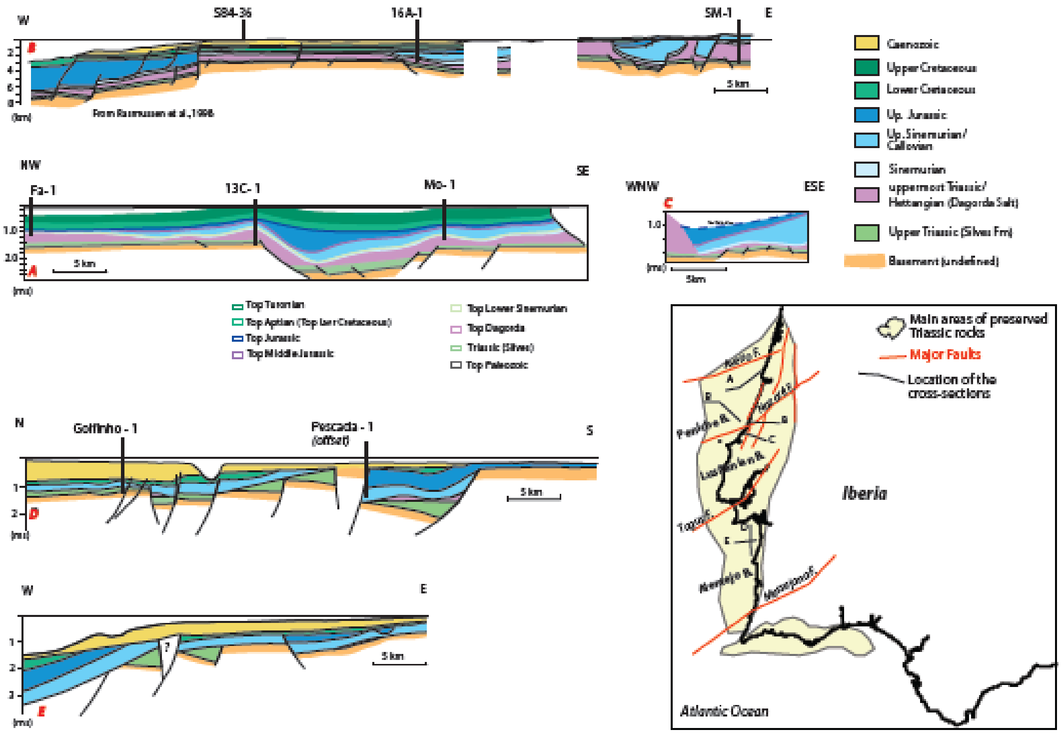

W
S84-36
16A-1
SM-1
E
B
(km)
From Rasmussen et al.,1998
5 km
Caenozoic
Upper Cretaceous
Lower Cretaceous
Up. Jurassic
Up. Sinemurian/
Callovian
Sinemurian
uppermost Triassic/
Hettangian (Dagorda Salt)
Upper Triassic (Silves Fm)
Basement (undefined)
NW
Fa-1
13C-1
Mo-1
SE
A
5 km
(ms)
WNW
ESE
C
5km
(ms)
Top Turonian
Top Aptian (Top Low Cretaceous)
Top Jurassic
Top Middle Jurassic
Top Lower Sinemurian
Top Dagorda
Triassic (Silves)
Top Paleozoic
Main areas of preserved
Triassic rocks
Major Faults
Location of the
cross-sections
Iberia
Atlantic Ocean
N
Golfinho - 1
Pescada - 1
(offset)
S
D
5 km
(ms)
W
E
?
E
5 km
(ms)

Figure 5A

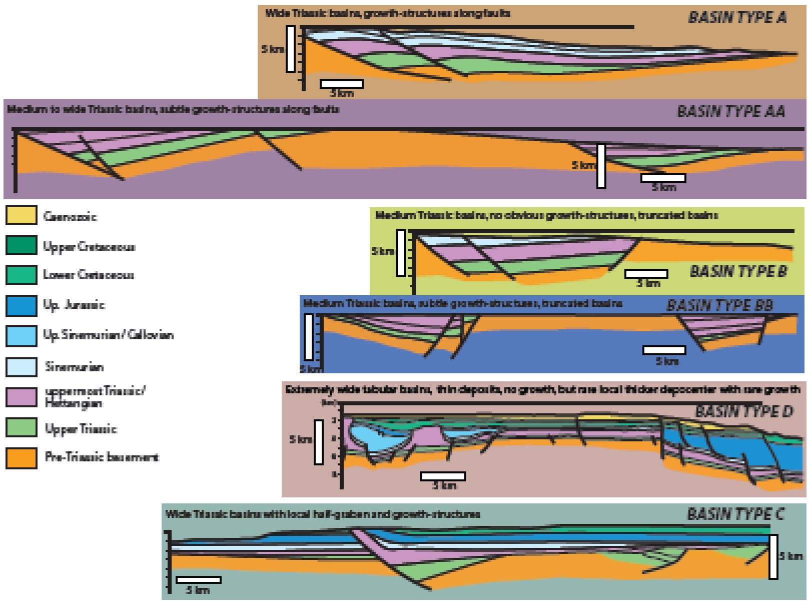

Wide Triassic basins, growth-structures along faults
BASIN TYPE A
5 km
5 km
Medium to wide Triassic basins, subtle growth-structures along faults
BASIN TYPE AA
5 km
5 km
Caenozoic
Upper Cretaceous
Lower Cretaceous
Up. Jurassic
Up. Sinemurian / Callovian
Sinemurian
uppermost Triassic/ Hettangian
Upper Triassic
Pre-Triassic basement
Medium Triassic basins, no obvious growth-structures, truncated basins
5 km
5 km
BASIN TYPE B
Medium Triassic basins, subtle growth-structures, truncated basins
BASIN TYPE BB
5 km
5 km
Extremely wide tabular basins, thin deposits, no growth, but rare local thicker depocenter with rare growth
BASIN TYPE D
5 km
5 km
Wide Triassic basins with local half-graben and growth-structures
BASIN TYPE C
5 km
5 km

Figure 5B

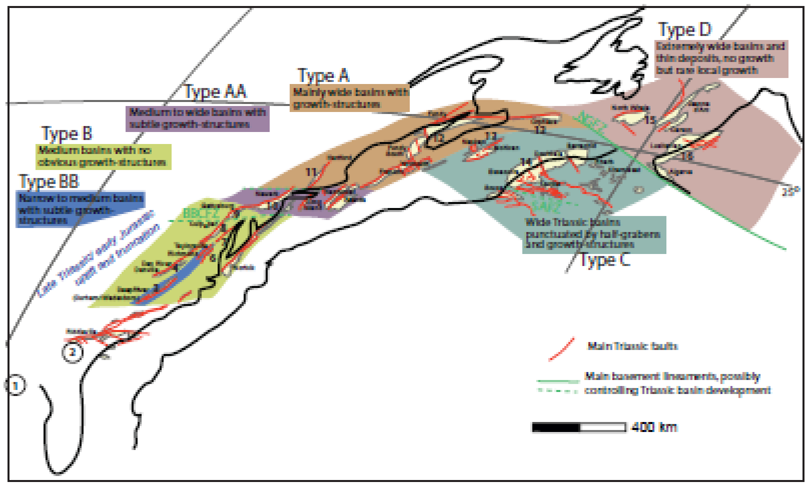

Type AA
Medium to wide basins with subtle growth-structures
Type A
Mainly wide basins with growth-structures
Type B
Medium basins with no obvious growth-structures
Type BB
Narrow to medium basins with subtle growth-structures
Late Triassic/ early Jurassic uplift and truncation
BBCFZ
Type D
Extremely wide basins and thin deposits, no growth but rare local growth
NGFZ
SAFZ
Wide Triassic basins punctuated by half-grabens and growth-structures
Type C
35°
Main Triassic faults
Main basement lineaments, possibly controlling Triassic basin development
400 km

Figure 6

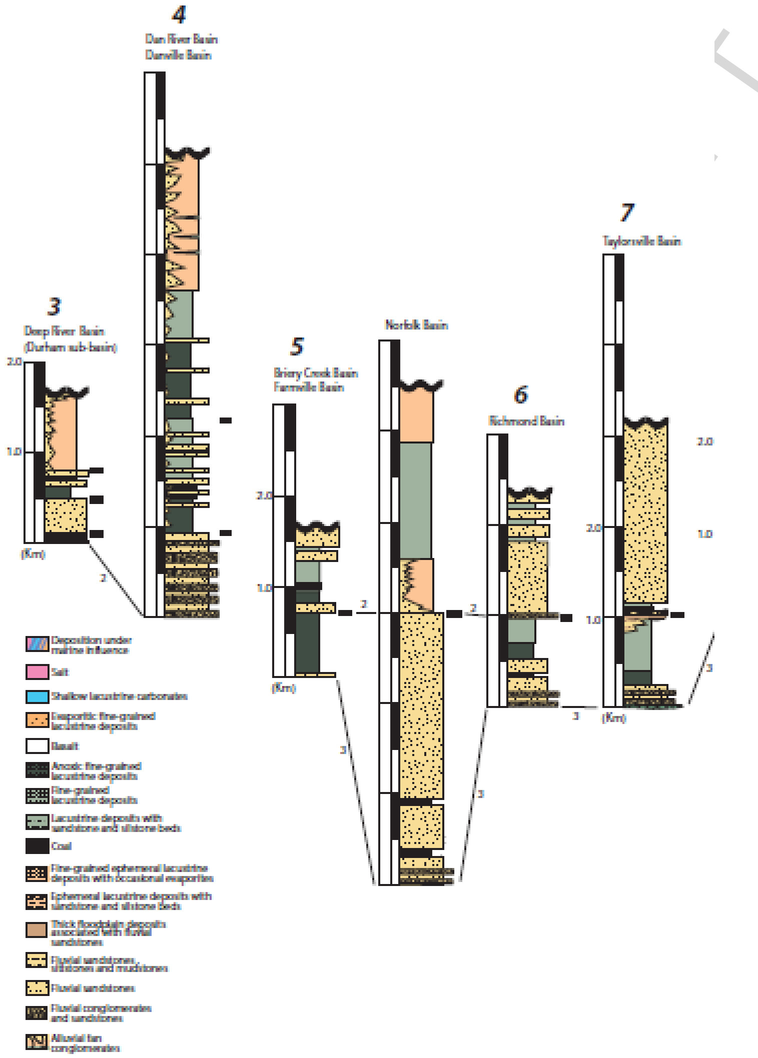

3
Deep River Basin
(Durham sub-basin)
4
Dan River Basin
Danville Basin
5
Briery Creek Basin
Farmville Basin
Norfolk Basin
6
Richmond Basin
7
Taylorsville Basin
2.0
1.0
(Km)
2
3
Deposition under marine influence
Salt
Shallow lacustrine carbonates
Evaporitic fine-grained lacustrine deposits
Basalt
Anoxic fine-grained lacustrine deposits
Fine-grained lacustrine deposits
Lacustrine deposits with sandstone and siltstone beds
Coal
Fine-grained ephemeral lacustrine deposits with occasional evaporites
Ephemeral lacustrine deposits with sandstone and siltstone beds
Thick floodplain deposits associated with fluvial sandstones
Fluvial sandstones, siltstones and mudstones
Fluvial sandstones
Fluvial conglomerates and sandstones
Alluvial fan conglomerates

Figure 6B

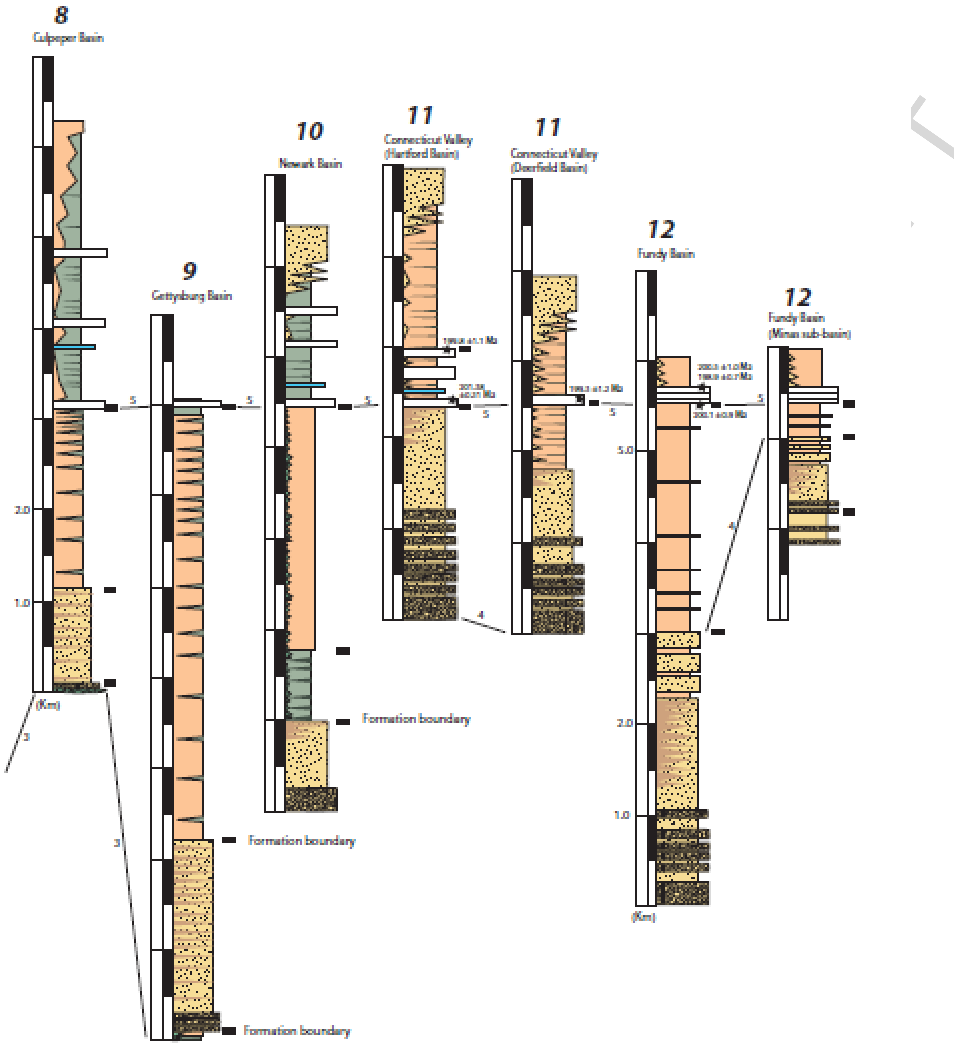

8
Culpeper Basin
9
Gettysburg Basin
10
Newark Basin
11
Connecticut Valley
(Hartford Basin)
11
Connecticut Valley
(Deerfield Basin)
12
Fundy Basin
12
Fundy Basin
(Minas sub-basin)
2.0
1.0
(Km)
5.0
Formation boundary
Formation boundary
Formation boundary

Figure 7

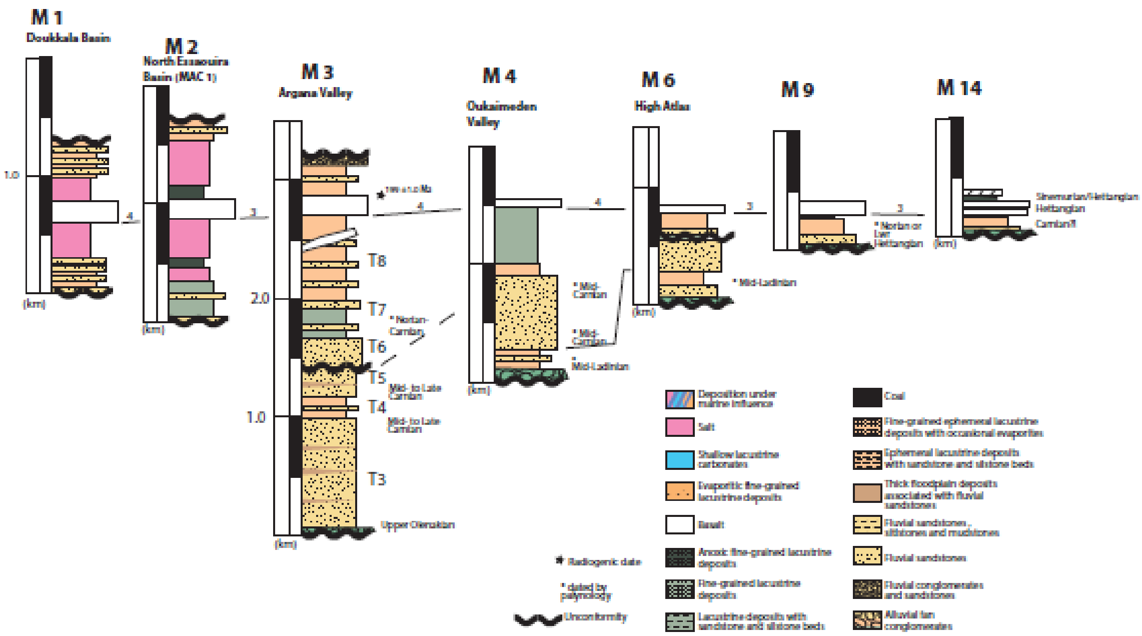

M 1
Doukkala Basin
M 2
North Essaouira Basin (MAC 1)
M 3
Argana Valley
M 4
Oukaimeden Valley
M 6
High Atlas
M 9
M 14
T8
T7
T6
T5
T4
T3
Upper Olenekian
* Mid-Ladinian
Sinemurian/Hettangian
Hettangian
Carnian?
Deposition under marine influence
Salt
Shallow lacustrine carbonates
Evaporitic fine-grained lacustrine deposits
Basalt
Anoxic fine-grained lacustrine deposits
Fine-grained lacustrine deposits
Lacustrine deposits with sandstone and siltstone beds
Coal
Fine-grained ephemeral lacustrine deposits with occasional evaporites
Ephemeral lacustrine deposits with sandstone and siltstone beds
Thick floodplain deposits associated with fluvial sandstones
Fluvial sandstones, siltstones and mudstones
Fluvial sandstones
Fluvial conglomerates and sandstones
Alluvial fan conglomerates
Radiogenic date
* dated by palynology
Unconformity

Figure 8

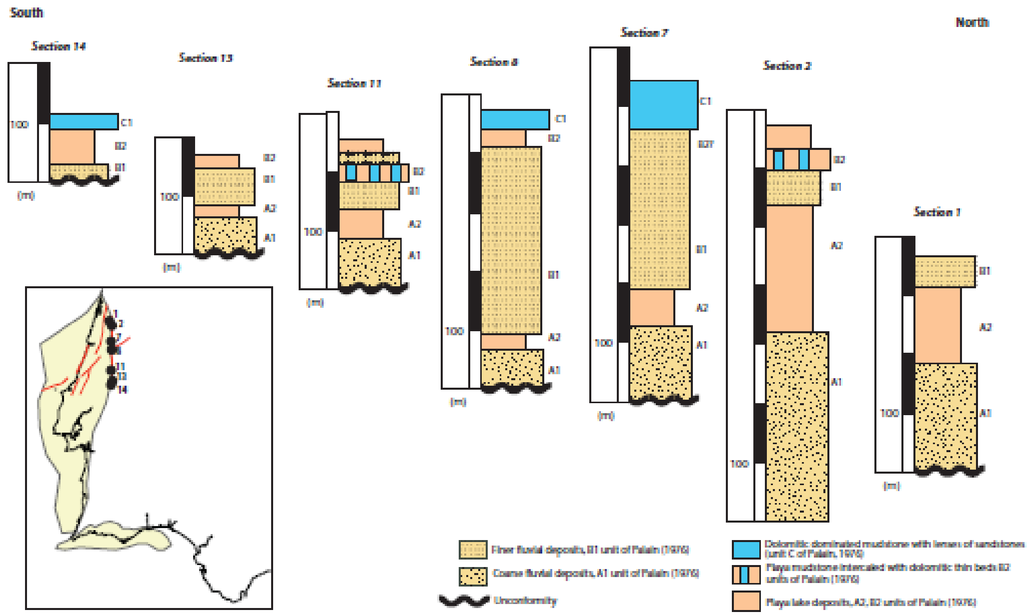

South
North
Section 14
Section 13
Section 11
Section 8
Section 7
Section 2
Section 1
100
(m)
C1
B2
B1
A2
A1
B2?
Finer fluvial deposits, B1 unit of Palain (1976)
Coarse fluvial deposits, A1 unit of Palain (1976)
Unconformity
Dolomitic dominated mudstone with lenses of sandstones (unit C of Palain, 1976)
Playa mudstone intercalated with dolomitic thin beds B2 units of Palain (1976)
Playa lake deposits, A2, B2 units of Palain (1976)

Figure 9

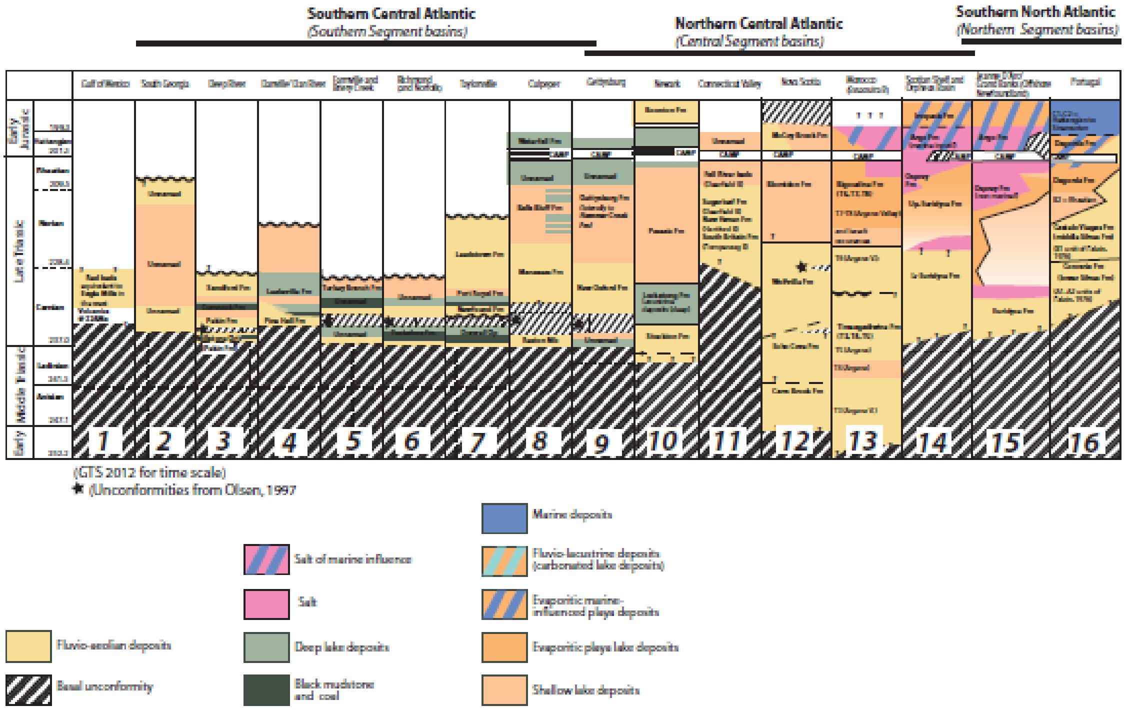

Southern Central Atlantic
(Southern Segment basins)
Northern Central Atlantic
(Central Segment basins)
Southern North Atlantic
(Northern Segment basins)
Gulf of Mexico
South Georgia
Deep River
Culpeper
Gettysburg
Newark
Connecticut Valley
Nova Scotia
Portugal
Early Jurassic
Late Triassic
Middle Triassic
Early
CAMP
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
(GTS 2012 for time scale)
(Unconformities from Olsen, 1997
Marine deposits
Salt of marine influence
Fluvio-lacustrine deposits
(carbonated lake deposits)
Salt
Evaporitic marine-influenced playa deposits
Fluvio-aeolian deposits
Deep lake deposits
Evaporitic playa lake deposits
Basal unconformity
Black mudstone and coal
Shallow lake deposits

Figure 10

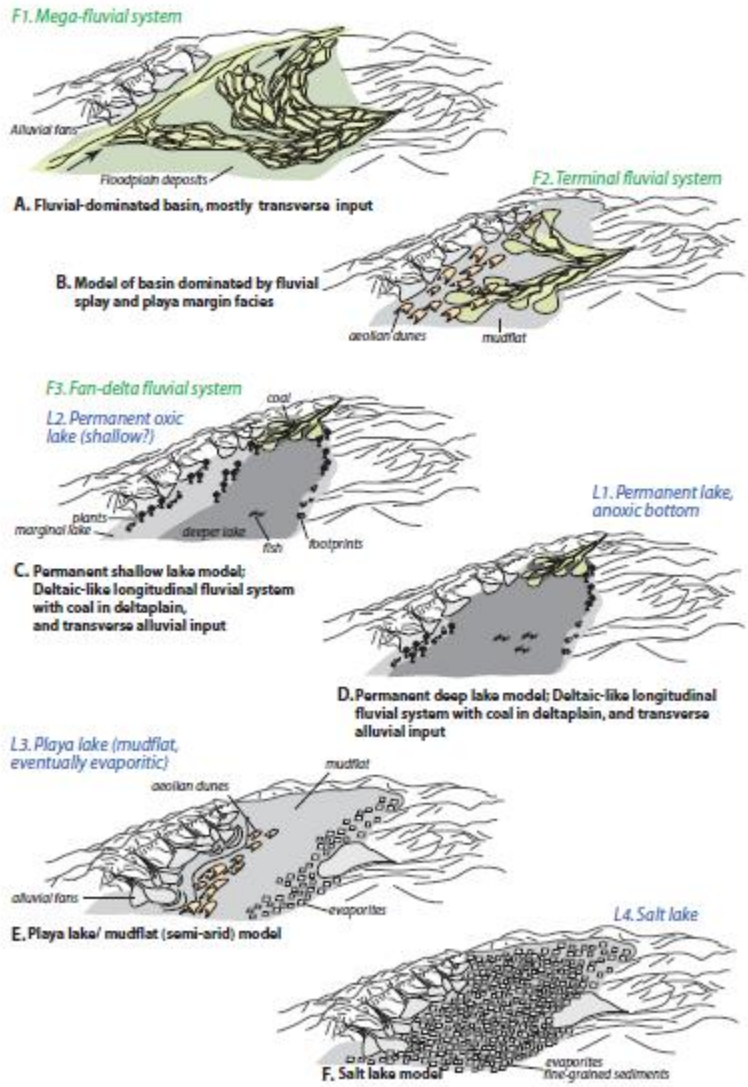

F1. Mega-fluvial system
Alluvial fans
Floodplain deposits
A. Fluvial-dominated basin, mostly transverse input
F2. Terminal fluvial system
B. Model of basin dominated by fluvial splay and playa margin facies
aeolian dunes
mudflat
F3. Fan-delta fluvial system
coal
L2. Permanent oxic lake (shallow?)
plants
marginal lake
deeper lake
fish
footprints
C. Permanent shallow lake model; Deltaic-like longitudinal fluvial system with coal in deltaplain, and transverse alluvial input
L1. Permanent lake, anoxic bottom
D. Permanent deep lake model; Deltaic-like longitudinal fluvial system with coal in deltaplain, and transverse alluvial input
L3. Playa lake (mudflat, eventually evaporitic)
mudflat
aeolian dunes
alluvial fans
evaporites
E. Playa lake/ mudflat (semi-arid) model
L4. Salt lake
F. Salt lake model
evaporites
fine-grained sediments

Figure 11A

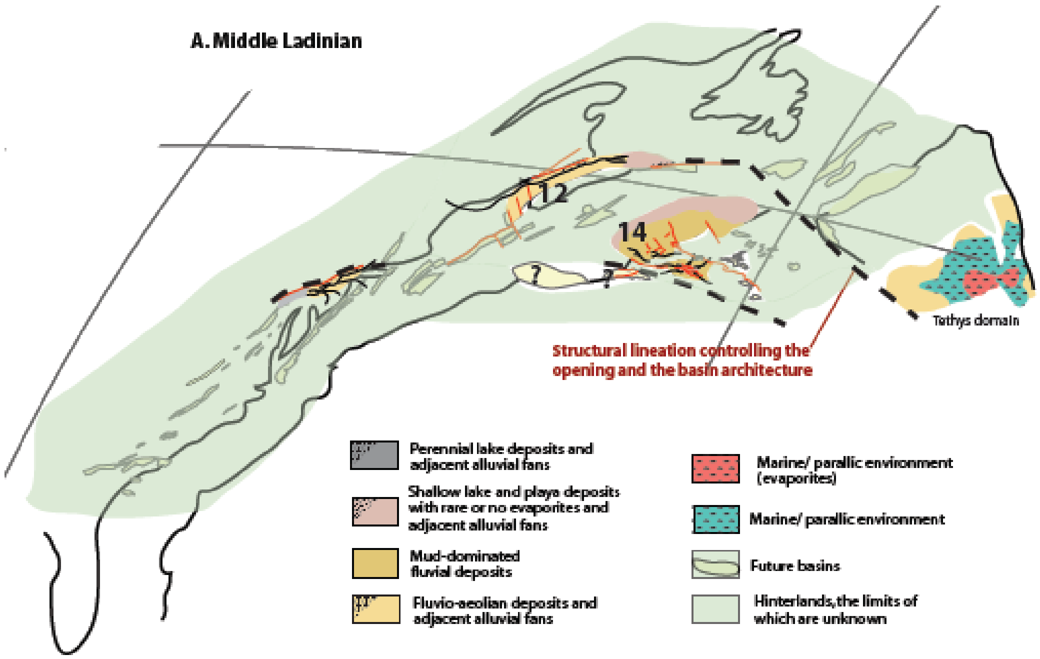

A. Middle Ladinian
12
14
?
?
Tethys domain
Structural lineation controlling the opening and the basin architecture
Perennial lake deposits and adjacent alluvial fans
Shallow lake and playa deposits with rare or no evaporites and adjacent alluvial fans
Mud-dominated fluvial deposits
Fluvio-aeolian deposits and adjacent alluvial fans
Marine/ parallic environment (evaporites)
Marine/ parallic environment
Future basins
Hinterlands, the limits of which are unknown

Figure11 B

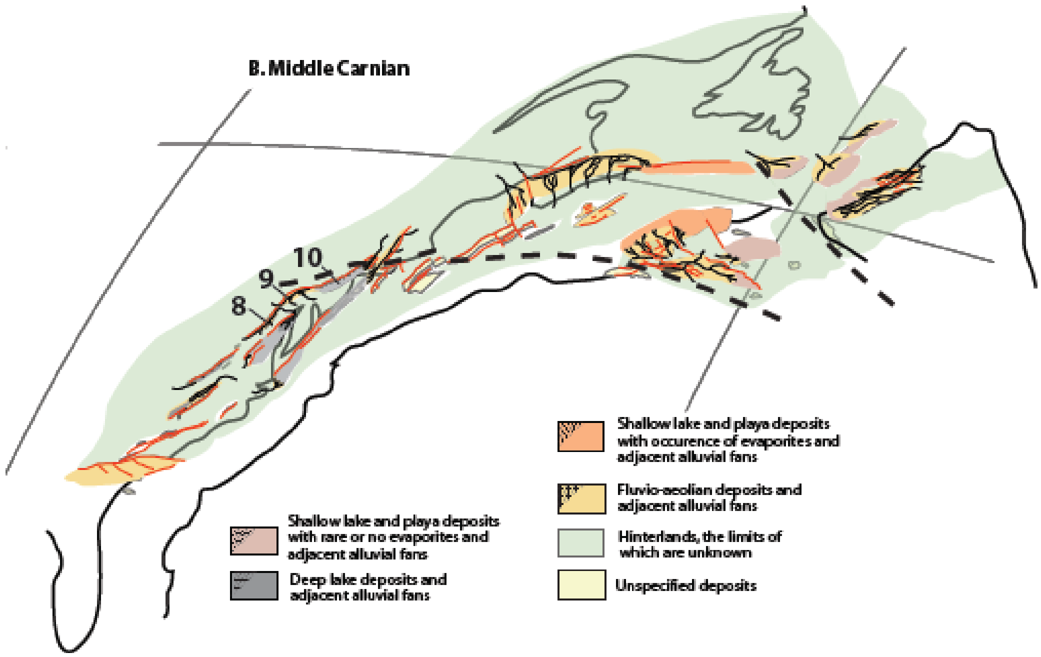

B. Middle Carnian
10
9
8
Shallow lake and playa deposits with occurence of evaporites and adjacent alluvial fans
Fluvio-aeolian deposits and adjacent alluvial fans
Shallow lake and playa deposits with rare or no evaporites and adjacent alluvial fans
Hinterlands, the limits of which are unknown
Deep lake deposits and adjacent alluvial fans
Unspecified deposits

Figure 11C

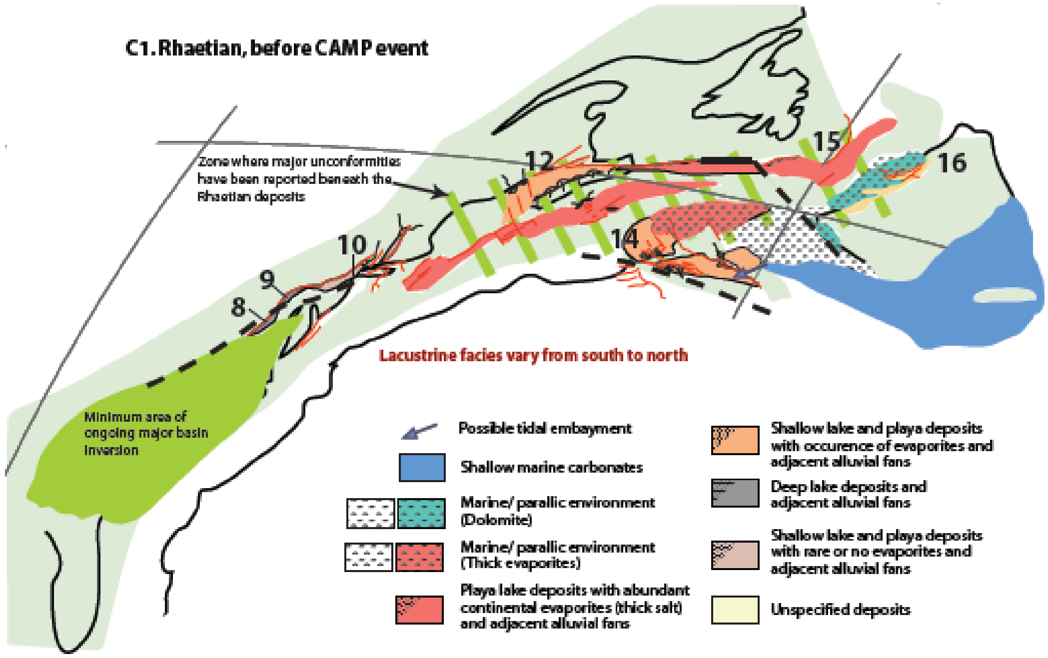

C1. Rhaetian, before CAMP event
Zone where major unconformities have been reported beneath the Rhaetian deposits
8
9
10
12
14
15
16
Lacustrine facies vary from south to north
Minimum area of ongoing major basin inversion
Possible tidal embayment
Shallow marine carbonates
Marine/ parallic environment (Dolomite)
Marine/ parallic environment (Thick evaporites)
Playa lake deposits with abundant continental evaporites (thick salt) and adjacent alluvial fans
Shallow lake and playa deposits with occurence of evaporites and adjacent alluvial fans
Deep lake deposits and adjacent alluvial fans
Shallow lake and playa deposits with rare or no evaporites and adjacent alluvial fans
Unspecified deposits

Figure 11CC

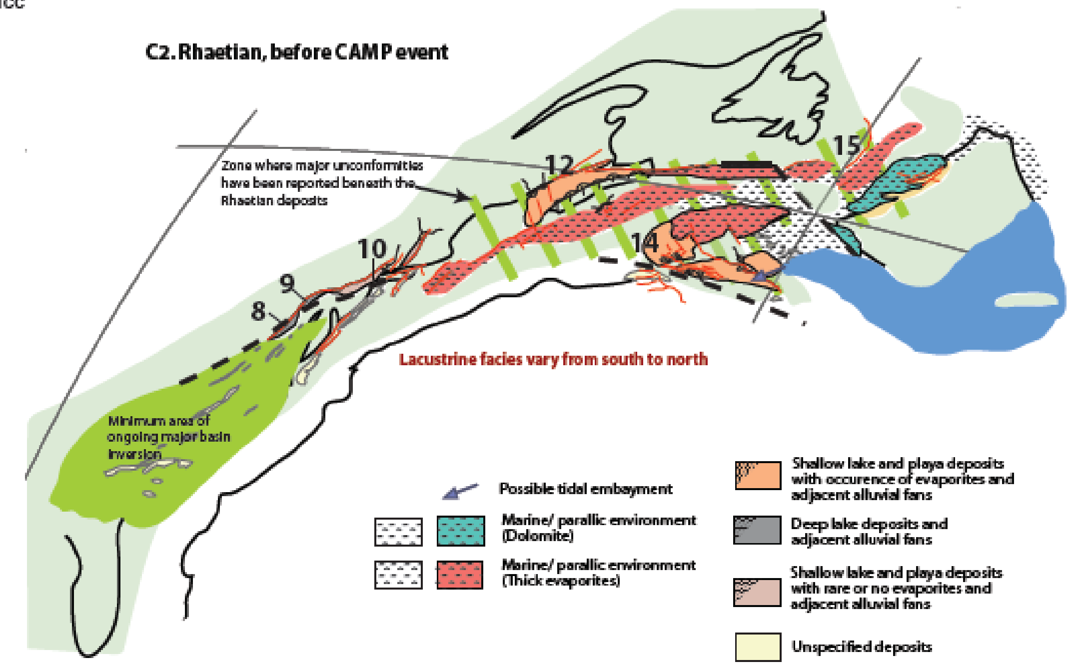

C2. Rhaetian, before CAMP event
Zone where major unconformities have been reported beneath the Rhaetian deposits
12
15
10
14
9
8
Lacustrine facies vary from south to north
Minimum area of ongoing major basin inversion
Possible tidal embayment
Marine/ parallic environment (Dolomite)
Marine/ parallic environment (Thick evaporites)
Shallow lake and playa deposits with occurence of evaporites and adjacent alluvial fans
Deep lake deposits and adjacent alluvial fans
Shallow lake and playa deposits with rare or no evaporites and adjacent alluvial fans
Unspecified deposits

Figure 11D

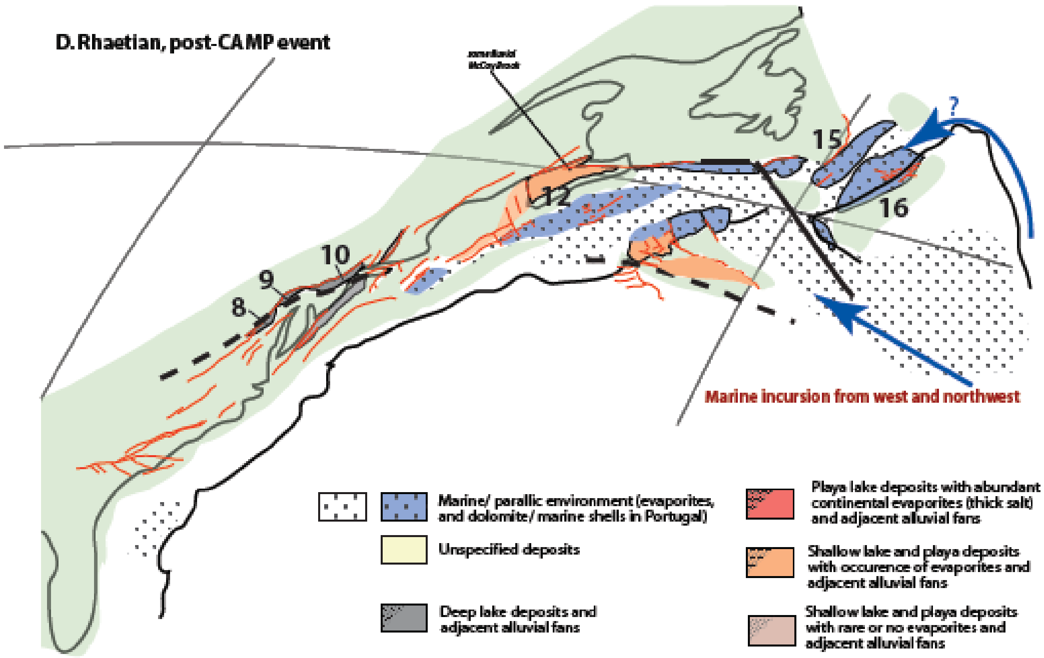
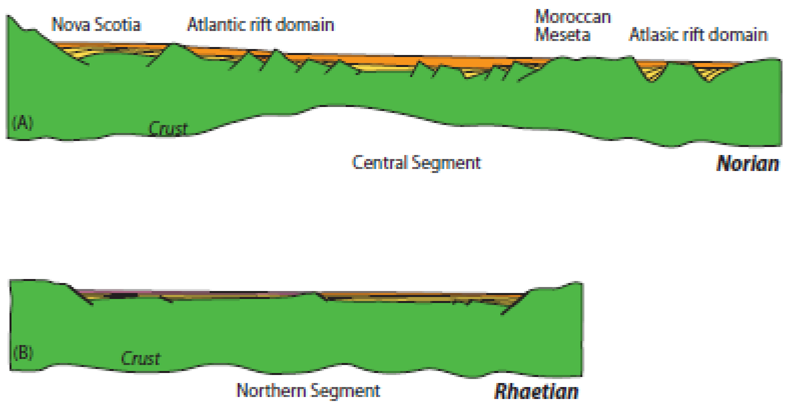

D. Rhaetian, post-CAMP event
8
9
10
12
15
16
?
Marine incursion from west and northwest
Marine/ parallic environment (evaporites, and dolomite/ marine shells in Portugal)
Unspecified deposits
Deep lake deposits and adjacent alluvial fans
Playa lake deposits with abundant continental evaporites (thick salt) and adjacent alluvial fans
Shallow lake and playa deposits with occurence of evaporites and adjacent alluvial fans
Shallow lake and playa deposits with rare or no evaporites and adjacent alluvial fans

Figure 12



Nova Scotia
Atlantic rift domain
Moroccan Meseta
Atlasic rift domain
(A)
Crust
Central Segment
Norian
(B)
Crust
Northern Segment
Rhaetian

Figure 13

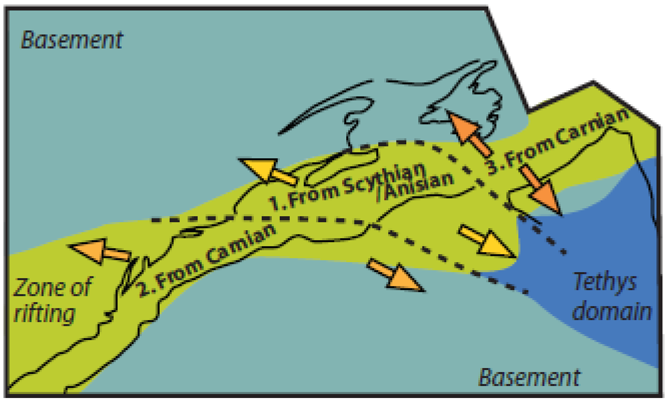
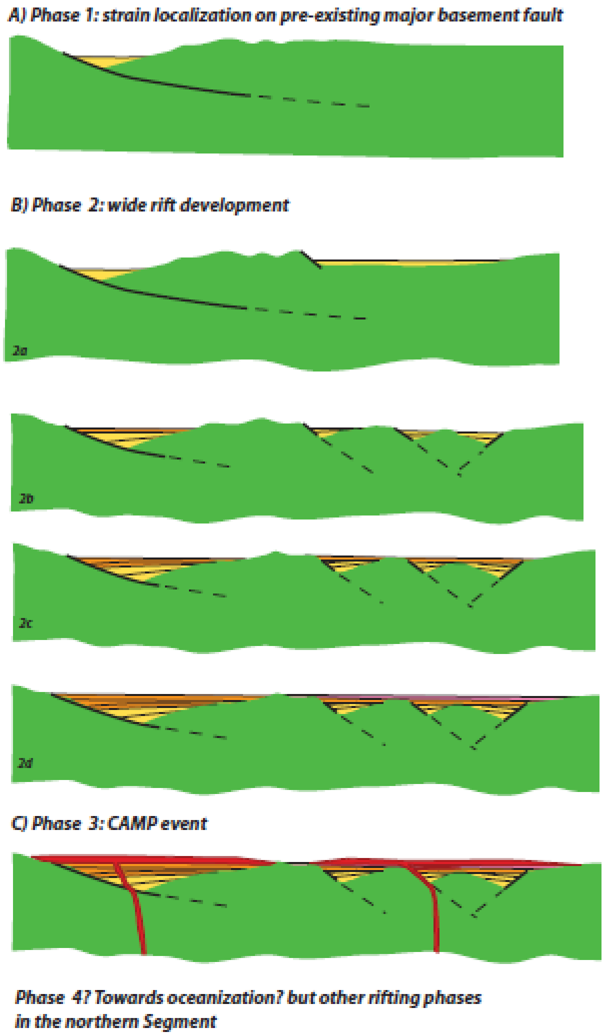

Basement
1. From Scythian /Anisian
3. From Carnian
2. From Camian
Zone of rifting
Tethys domain
Basement

Figure 14


A) Phase 1: strain localization on pre-existing major basement fault
B) Phase 2: wide rift development
2a
2b
2c
2d
C) Phase 3: CAMP event
Phase 4? Towards oceanization? but other rifting phases
in the northern Segment

North American basins facies	Brief description (From Smoot, 1991, modified)	Interpretation (Smoot, 1991)	More details and interpretations	Main depositional environments
(AF1)	Matrix-supported conglomerates poorly sorted, comprising boulders, and forming lenses convex-upwards	debris flow deposits	debris flow	**Alluvial-fan deposits**
(AF2)	Clast-supported conglomerates forming lenses locally convex-upwards, and interbedded with sandstone beds	deposited by flash-flooding shallow-braided streams	debris flow or hyperconcentrated flow	
(AF3)	Conglomerates presenting cross-bedding and imbrications	formed by braided stream processes on fans	stream flow processes	
(AF4)	Pebbly muddy sandstones forming isolated channel-form lenses and comprising imbricated cobble, cross-beds and horizontal stratification,	deposits of hyperconcentrated sheetfloods or braided stream deposits	stream flow processes	
(F1)	Imbricate boulder and cobble conglomerate, poorly sorted pebbly sandstone and sandstones presenting crude upwards-fining sequences (0.3 – 2.0 m thick). Sandstones are dominated by planar horizontal or low-angle lamination	braided stream deposits		**Coarse fluvial deposits**
(F2)	Poorly sorted cross-bedded pebbly sandstones overlying cobble and boulder conglomerate and presenting fining-upwards sequences (2.0 – 6.0 m thick), abruptly overlain by heavily bioturbated mudstone or siltstone.	It is interpreted to be deposited by braided streams, locally anastamosing on vegetated muddy plains	It seems that F1 and F2 are lateral equivalent of a similar system	
(F3)	(F3) is similar to (F2) but conglomerates are moderately sorted	Braided stream	F1, F2, F3 were deposited by the same fluvial system	
(F4)	Sandstone, silstone, and mudstone forming rhythmically stacked upwards fining-sequences (2-7 m thick). Sandstones	This facies is interpreted as lateral accretion on point bar and therefore as a		**Sandy fluvial deposits**

	comprise trough cross-beds at the base grading to ripple cross-laminated fine sandstone capped by heavily bioturbated. Stacked sequences occur as a depositionally inclined lense that is thicker and coarser downdip	meandering fluvial system		
(F5)	(F5) is similar to (F4) but contains mud lenses and coarse scour at the base of sandstones.	Meandering fluvial system	(F4) and (F5) should be grouped in a single facies	
(F6)	Mudstone, siltstone and fine-grained sandstone with abundant root structures, burrows, and locally carbonate nodules	interpreted as fluvial flood plain where soil development occurred		**Floodplain deposits**
(F7)	Poorly sorted conglomerate with angular clasts forming imbrications and cross-beds in places.	This is interpreted as colluvium deposited within bedrock topography.	Alluvial fan deposits	**Colluvium deposits**
(L1)	Finely laminated, organic-rich shale and limestone alternating with silstone beds. Burrows are locally abundant, fossil-fish well preserved and aquatic reptile and conchostracans and ostracods abundant.	This facies is interpreted to be deposited in deep perennial lakes		**Perennial lake deposits**
(L2)	Thin-bedded to massive mudstone, commonly organic-rich with abundant burrows, alternating with graded sandstones presenting soft-sedimentation deformation. Ostracods, conchostracans and wood fragment are common. Polygonal cracks and root structure are occasional.	This facies is interpreted to be deposited in shallow perennial lake, or margin of deeper lake with occasional subaerial exposure		
(L3)	Thin-bedded mudstone and siltstone with abundant polygonal cracks. Siltstone and occasional sandstone beds present abundant scours and intraclasts.	This facies was deposited in shallow ephemeral lake or fluctuating margin of larger lake		**Ephemeral lake deposits**
(L4)	Massive mudstone with	This facies		

	abundant, sinuous polygonal cracks. Breccia-like fabric of mudstone separated by silt-rich cracks fillings gradationally overlies L2 or L3 deposits.	represents a period of playa dry mud-flat showing alteration of wet and dry periods		
(L5)	Poorly sorted, sandy mudstone with irregular pods of siltstone and sandstone and irregular sandstone beds with hump-shape ripples and deformed boundaries.	This facies represents salt-encrusted saline mudflat		**Saline mudflat**
(L6)	Massive mudstone and siltstone with abundant cement- or sediment-filled root structures, carbonate nodules, remnant patches of rippled sandstone, and polygonal cracks.	This facies represents vegetated mudflat		**Mudflat lateral to playa lakes or floodplain with crevasse splay deposits**
(LM1)	Channel-form sandstone or conglomerate lenses (10 to several 10's meter wide and 2-10 m thick) intercalated with (L1) or (L2) facies. Sandstone lenses present trough cross-bedding and grade upwards to ripple cross-lamination with soft-deformation structures. Bioturbation may disrupt the top of sequences.	This facies represents delta deposits when lake level drops	Fluvial channel deposits eroding the playa during a lake level fall	**Distal fluvial channel deposits of a terminal splay, eroding playa during a lake level fall**
(LM2)	Depositionally inclined sandstone sets stacked to form upward coarsening sequences (10-40 m thick). The low-angle sets in the lower portion of the sequence are dominated by climbing ripples.	This facies has interpreted as deposits of Gilbert-type delta	Fluvial terminal splay deposits on the playa margin or fan delta deposits if water level is relatively high	**Terminal splay deposits or fan delta deposits depending on lacustrine level (depending on associated lacustrine facies)**
(LM3)	Wedge-shape sandstone sheets comprising sandstone thin beds of climbing ripples intercalated by clay partings with polygonal cracks.	This facies represent sheet-like deltaic plain deposits produced by the intersection of very shallow lake with flash-	This facies is equivalent to terminal splay that in the terminology used in Leleu & Hartley 2010	**Distal terminal splay deposits**

		flooding streams.		
(LM4)	Graded sandstones to conglomerates forming coarsening-upward sequences (3-10 m thick) from (L1) or (L2) deposits. Lower sequence dominated by oscillatory ripple cross-lamination grading up to flat-bedded pebbly sandstones. Upper sequence is poorly sorted boulder to cobble conglomerates comprising patchy rippled-sandstones.	This facies is interpreted as wave sorted shoreline deposits	Alluvial-fan deposits on the margin of the lake	**Fan-delta deposits.**
(LM5)	Similar facies than LM4 but conglomerates absent.	It is interpreted to be a distal equivalent to (LM4)	Alluvial-fan deposits on the margin of the lake	

Table 1: Lithofacies description and interpretation of North American Triassic basins (from Smoot, 1991, modified) and re-interpretation in term of main palaeo-environments.

Essaouira Basin (Morocco)	**Brief description**	**Interpretation**	**Depositional environment**
T3 Member (Tanameurt Conglomerate)	Conglomerates that comprise abundant angular volcanic rocks fragments (25%) as well as pebbles of quartzite, limestone, phyllite, siltstone and sandstone. The clasts are aligned along large-scale trough and low angle planar cross-beds	Fluvial deposits (Jones, 1975)	**Fluvial or alluvial fan?**
T4 Member (Aglelgal Sandstone)	It comprises alternating coarse-grained sandstone units and thick intervals of mudstones interbedded with siltstone and fine-grained sandstones. Coarse-grained units comprise immature lithic sand and display channel-like geometries. The mudstone intervals are characterized by tabular bedding and occasionally wedge-shape beds in accretion sets with desiccation cracks, rooting structures, caliche horizons and footprints.	Brown (1980) interpreted T4 environment as a braid-plain of high sinuosity meandering rivers but Hofmann et al. (2000) interpret those deposits as playa and sheetflood deposits.	**Playa margin and distal splays**
T5 Member (Irohalene Mudstone)	The basal part of T5 is dominated by mudstones in which bedding is continuous and parallel. Silt and fine-grained sand grains are present in the mudstone which does not show any grading and in which depositional structures are not discernable but mottling is abundant. The upper part of T5 contains sandstone beds which get thicker upwards. The uppermost sandstone beds (6-10 m thick) are intercalated with mudstones but laterally they amalgamate and form a 90 m thick unit. Ripples and wavy lamination pass into low-angle or continuous parallel cross-beds upwards. No evidence of subaerial exposure is found. Estheriids were found by Brown (1980) in the top of the T5 succession and thought to be indicating brackish water conditions.	More recent interpretations attribute T5 deposits to meandering ephemeral streams (Tourani, unpublished data). The boundary between T4 and T5 is debated (Tixeront, 1973; Brown, 1980; Hofmann et al., 2000) but the facies variation is probably gradual.	**Fluvial and distal splays deposits**
T6 Member (Tadrart Ouadou Sandstone)	T6 have been mapped as “discordant” on the underlying strata by Tixeront (1971) and “unconformable” by Olsen (1997) but Brown (1980) suggested that it was concordant. It is still debated. Hofmann et al (2000) describe it as conformable on T5 but locally disconformable. T6 comprises three distinct channel-like bodies located at different places in the Argana Valley. Brown (1980) also shows that when sandstone bodies pinch out, mudstones are lateral equivalent. He suggested that the overlying Sidi Mansour	Proximal braided river conglamerates grading upwards into sand-dominated distal braided river deposits intercalated with aleolian sandstones (Tourani, unpub., Hofmann et al., 2000)	**Coarse fluvial grading to sandy fluvial deposits**

	Mudstone (T7 Member; 0-200 m in thickness) was partly contemporaneous with T6. However Hofmann et al. (2000) describe a sharp contact between T6 and T7. The sandstones consist of mature sand with bed thicknesses of 1-5 m, with thicker beds occurring towards the top of sand bodies. They contain parallel and low-angle cross-bedding, climbing ripples and convolute structures. In the southern body, thin halite beds are interbedded with sandstone and siltstone beds.		
T7 Member (Sidi Mansour Mudstone)	T7 member comprises graded sandstone beds at the base and evolves to interbedded rippled siltstone and mudstone beds which get thicker upwards where mottling is frequent. It is a fining-upwards sequence of sandstone, siltstone and brown mudstone. The maximum thickness is in the centre of the Argana Valley. It contains three reduced grey, green and white marly shales containing copper mineralization, plant debris, and locally Estheriids (Defretin & Fauvelet, 1951) though to be indicator of brackish conditions.	Brown (1980) interpreted T7 as deposits of delta plain and delta front environments. Hofmann et al. (2000) suggested that T7 and T8 are similar and comprise cyclically-arranged mud-rich facies and cannot be separated into two distinct lithofacies units and interpret those deposits as part of a playa system.	**Playa deposits**
T8 Member (Hasseine Mudstone)	T8 was described by Brown (1980) as containing mainly of claystones, siltstones and very fine sandstones with minor amounts of chlorite, halite, gypsum and anhydrite. Mudstones and siltstones are extensive and tabular, and are interbedded with fine sandstones forming wide (< 50 m) thin (< 0.5 m) lenses.	T8 is highly bioturbated and was intepreted by Brown (1980) to represent tidal flat deposits. It is now accepted that the T7/ T8 deposits were formed in shallow ephemeral lakes, saline mudflats with periodic fluvial and aeolian inputs (Hofmann et al., 2000).	**Playa deposits**

Table 2: Description of sedimentary Members of the Essaouira Basin (Morocco), interpretations from various authors, and re-interpretation in term of main palaeo-environments.